Geol Survey

STATE OF ILLINOIS
DEPARTMENT OF REGISTRATION AND EDUCATION

Temperature Prospecting for Shallow Glacial and Alluvial Aquifers in Illinois

Keros Cartwright

ILLINOIS STATE GEOLOGICAL SURVEY
John C. Frye, *Chief* URBANA

CIRCULAR 433 1968

TEMPERATURE PROSPECTING FOR SHALLOW GLACIAL AND ALLUVIAL AQUIFERS IN ILLINOIS

Keros Cartwright

ABSTRACT

Theoretical considerations of the thermal properties of glacial and alluvial deposits in Illinois suggest that a shallow aquifer might form a heat sink (or source) that would influence the temperature effects on the soil of heat originating at the land surface and within the crust. If shallow aquifers are nonuniformly distributed laterally, and if their effects on the temperatures of surface soil can be measured and distinguished from other factors that affect soil temperature, a possible exploration method for shallow aquifers is suggested. A positive (warm) anomaly would be expected over such aquifers in the winter and a negative anomaly in the summer. The size of the anomalies is dependent upon the thermal properties of the overburden, temperature difference between the surface and aquifer, and the depth of burial of the aquifer.

An electronic thermometer, utilizing a thermister at the end of an aluminum-tipped stake and a transistor-amplified bridge circuit, was used to measure soil temperatures at a depth of about 18 inches in seven areas in Illinois where shallow aquifers were known in some detail. The data from these surveys, presented here, show maximum anomalies of about 2° C over the aquifer. Surveys were made both in summer and winter; summer anomalies generally are of a greater magnitude than winter anomalies.

Soil differences, vegetation, and ice in the soil also affect the soil temperatures. The effect of vegetation in the summer can be as great as the anomaly produced by an aquifer. Frozen soil tends to eliminate anomalies, but this effect can be partly overcome by taking readings at a greater depth. In general, however, field data show close agreement between the location of shallow aquifers and thermal anomalies.

INTRODUCTION

Most geophysical techniques used in the exploration for ground water measure some property of the rocks, rather than properties of water. However, the presence of water in the rocks affects the results somewhat in electrical earth resistivity and seismic methods, the most commonly used exploration techniques. A property that may be exploited in ground-water exploration is the high specific heat of water or its resistance to changes in temperature.

Theoretical considerations of glacial and alluvial deposits suggest that a saturated aquifer may disturb the geothermal gradient by acting as a heat sink (absorbing heat) (Lovering and Goode, 1963) or heat source. This disturbance may influence the surface soil temperature. If surface soil temperature variations resulting from disturbance of the geothermal gradient by a shallow aquifer can be measured, the presence of the aquifer might be detected, provided the aquifer has lateral boundaries across which will be temperature contrasts and provided the temperature effects of other heat sources (or losses) can be eliminated or evaluated. The glacial drift of Illinois contains many shallow linear deposits; temperature prospecting over several of them has suggested that the method bears further investigation.

Nomenclature

The following symbols are used throughout the paper in equations, figures, and tables:

A Area—cm^2

α Thermal diffusivity—cm^2/sec

β Variable of integration, in this case $\beta = \frac{x}{2\sqrt{t\alpha}}$

c Specific heat—cal/(gm)(°C)

k Thermal conductivity—cal/(sec) (cm) (°C)

ℓ Lower boundary of the slab (overburden) above the heat source

m Upper boundary of the slab above the heat source

n Degrees of freedom—$n = 1$ for one dimension, $n = 2$ for two dimensions, and $n = 3$ for three dimensions of heat flow

Q Quantity of heat—cal

Q' Rate of heat production in a permanent heat source—cal/sec $\cancel{ft^2}$ cm^2 for one dimension

ρ Density—gm/cm^3

S' Strength of heat source

t Time—seconds

T Temperature—°C

T_r Surface temperature range/2

x Depth variable—cm

B Constant of integration

C Constant of integration.

AQUIFER TEMPERATURES

Geothermal Gradients

Bedrock

The exact geothermal gradients in Illinois are not known, but they can be expected to vary from one area to another. A commonly accepted value is 1° F increase in temperature per 100 feet of depth (18.23° C per kilometer).

Suter et al. (1959) reported a gradient of 1° F per 100 feet in the deep aquifers in northern Illinois and only a slight gradient in the shallow aquifers. The shallow aquifers—glacial drift and shallow dolomite—contained water only slightly warmer than the mean annual air temperatures. These temperatures were obtained mainly from wells in aquifers from which large amounts of water are often pumped.

Estimates of geothermal gradient in the Illinois Basin vary slightly. Pryor (1956) used 1° F per 100 feet as the gradient in the Illinois Basin. McGinnis (1968), using a large number of data from the oil fields, found the gradient to be about 1.1° F per 100 feet (20.05° C per kilometer) with the gradient increasing slightly in the shallower rocks. Loofbourow (1966) suggested an average gradient of 2° F per 100 feet (36.46° C per kilometer) in the oil-producing area of the basin.

The geothermal gradient varies from place to place, depending upon rock type, age, and moisture content. In general, the older and more compact the rock, the lower the geothermal gradient; thus, gradients around 0.5° F per 100 feet (9.12° C per kilometer) are recorded for the Canadian Shield, whereas gradients of 2° F per 100 feet (36.46° C per kilometer) are common in the Mississippi Embayment of Louisiana (Loofbourow, 1966; Spicer, 1942).

Variations in temperature gradients are best explained by considering the nature of heat flow. The quantity of heat (Q) that flows to the surface per unit area depends upon the thermal conductivity (k) of the rocks and the thermal gradient ($\delta T/\delta x$):

$$Q = -k(\delta T/\delta x) \tag{1}$$

or for a constant gradient through the vertical interval $x_1 - x_2$ and surface area A:

$$Q = kA \frac{(T_2 - T_1)}{(x_1 - x_2)} \tag{2}$$

where $T_2 > T_1$. Under steady-state conditions, the quantity of heat transmitted vertically across any unit thickness of rock is constant. The temperature gradient within each rock unit will vary inversely with its thermal conductivity:

$$\frac{k_1}{k_2} = \frac{(\delta T_2/\delta x_2)}{(\delta T_1/\delta x_1)} \,. \tag{3}$$

Equation (3) can be applied to the problem of gradients in the Illinois Basin. The section of rocks from the top of the Cambrian Mt. Simon Sandstone megagroup to the top of the Mississippian Valmeyeran Series is in large part limestone and dolomite with lesser amounts of shale and sandstone. Above this section, the Mississippian Chesterian and the Pennsylvanian rocks are predominantly clastic, with shale as the major rock type. The thermal conductivities of individual rocks vary. The thermal conductivities reported (presumably water saturated) for limestone and dolomite (Lovering and Goode, 1963; Loofbourow, 1966; Spicer, 1942; Handbook of Chemistry and Physics, 1967) range from 0.0048 to 0.0115 cgs units, with most in excess of 0.0065; sandstones have thermal conductivities of about 0.0055 and shale 0.0042 to 0.0052. Assuming a geothermal gradient of 18.23° C per kilometer (1° F per 100 feet), a thermal conductivity of 0.007 in the predominantly carbonate section of the Illinois Basin, and a thermal conductivity of 0.005 in the predominantly shale section, a gradient of 25.52° C per kilometer (1.4° F per 100 feet) is obtained for the shale. This probably is a reasonable figure because when the two gradients are combined, a geothermal gradient of about 20.96° to 21.88° C per kilometer (1.15° to 1.2° F per 100 feet), depending on the ratio of the thicknesses assumed, is obtained for the rocks of the Illinois Basin.

Glacial Drift

Measurements of geothermal gradients have not been made in glacial drift and, therefore, the geothermal gradients must be estimated from theoretical considerations. Observations of water temperature, generally measured in pumping wells, have been made in a number of areas in the state. In the Chicago region, Suter et al. (1959) reported that the temperature of water from 213 drift and shallow dolomite wells ranged from 46° to 54° F (7.8° to 12.3° C), averaging 51.6° F (11.0° C), with 71 percent of the temperatures between 50.5° and 52.5° F (10.4° and 11.5° C). The mean annual air temperature of the Chicago region ranges from 48° to 51° F (9.0° to 10.6° C).

Walker, Bergstrom, and Walton (1965) reported water temperatures ranging between 53° and 57° F (11.8° and 14.0° C) and averaging 55° F (12.9° C) in the Havana region of Illinois, in areas unaffected by river infiltration. The mean annual air temperature of this region is about 51° F (10.6° C).

A tabulation of the data presented by Hanson (1950, 1958, 1961) for municipal water supplies in the east-central region of Illinois shows the water temperatures from wells 50 to 400 feet deep to range from 53.5° to 55.5° F (12.0° to 13.2° C), averaging 54.5° F (12.6° C). The approximate mean annual air temperature at Champaign-Urbana is 52° F (11.2° C). A linear regression line drawn through the data for east-central Illinois (fig. 1) suggests a gradient of 0.15° F per 100 feet (2.73° C per kilometer), although an argument could be made for a higher gradient by ignoring the points less than 100 feet and greater than 300 feet deep. However,

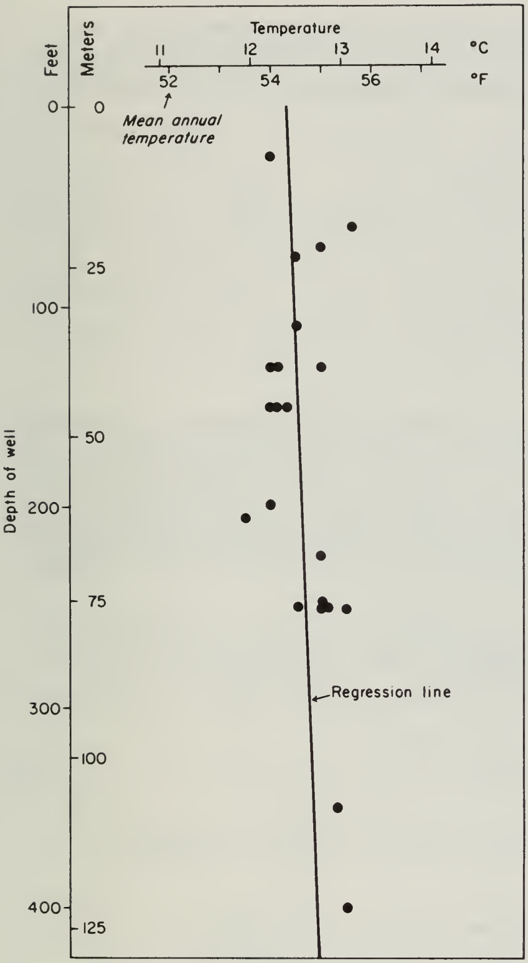

Figure 1 - Water temperatures from pumping wells in east-central Illinois (from Hanson, 1950, 1958, 1961).

all temperatures are for water pumped from water wells and suggest either that there is a low geothermal gradient (less than 18.23° C per kilometer, or 1° F per 100 feet) or that the aquifers form anomalous temperature bodies in the earth. The latter case is probably correct, as will be shown later. Also of significance is the fact that the water temperatures are about 2° F (1.1° C) above mean annual air temperatures.

If the geothermal gradient in the glacial drift is approached theoretically in the same manner that the discrepancies of reported geothermal gradients in the bedrock were resolved, the results are quite different from the results given by water temperatures. Bredehoeft and Papadopulos (1965) and Birch (1942, p. 259) used 0.002 cgs units as a typical value of the thermal conductivity of water-saturated clay. Loofbourow (1966), Misener, Thompson, and Uffen (1951), and Spicer (1942) give values of thermal conductivity ranging between 0.0021 and 0.0037 for moist to wet soils. Penrod, Elliott, and Brown (1960) give a value of 0.0017 for dry clay soil. Lovering and Goode (1963) found a value of 0.0024 in dry Quaternary gravel in Utah. A value of 0.0025 cgs units is assumed an average value for moist to wet glacial till in Illinois in this report.

Assuming values of 0.0025 for the thermal conductivity of glacial till, a thermal conductivity of 0.005 and a geothermal gradient of 25.52° C per kilometer (1.4° F per 100 feet) for the underlying Pennsylvanian bedrock, and substituting these values in equation (3), a geothermal gradient of 51.04° C per kilometer (2.8° F per 100 feet) is obtained for glacial till. If the value of thermal conductivity of 0.002 is used for glacial till, a geothermal gradient of 63.81° C per kilometer (3.5° F per 100 feet) is obtained, and if 0.003 is used, a gradient of 43.75° C per kilometer (2.4° F per 100 feet) is obtained. Or, if an average gradient for the whole basin is assumed to be 21.88° C per kilometer (1.2° F per 100 feet), with an average thermal conductivity of 0.006 for bedrock and a drift thermal conductivity of 0.0025, the geothermal gradient will be 52.50° C per kilometer (2.88° F per 100 feet). All of these calculations suggest a geothermal gradient of slightly less than 54.69° C per kilometer (3° F per 100 feet); a value of 51.04° C per kilometer (2.8° F per 100 feet) probably is a good average value in areas of clayey glacial till directly over Pennsylvanian bedrock.

The discrepancy between the theoretical value of geothermal gradient just obtained and the observed temperatures of water from shallow wells may be used to provide a clue to the presence of shallow glacial and alluvial aquifers.

SURFACE EFFECTS

Two obvious factors that affect shallow soil temperatures are the diurnal and seasonal temperature changes. These are both periodic variations, and their fluctuation can be approximated by a sinusoidal curve. A third factor affecting near surface temperatures is nonperiodic weather variations, such as warm or cold periods of long duration. The soil stores heat during warm periods and releases it during cold periods.

The depth to which changes of air temperature affect ground temperatures was investigated in detail by Lovering and Goode (1963) and Penrod, Elliott, and Brown (1960). Jaeger (1965) states that in most cases, surface effects are negligible below a depth of about 65 feet. Penrod, Elliott, and Brown (1960) gave this depth as 100 feet. Lovering and Goode (1963) estimated this depth to be between 30 and 130 feet, depending upon the thermal diffusivity constant and the duration and magnitude of the surface-temperature fluctuations.

The thermal diffusivity constant (α) equals the rise in temperature per unit volume produced by a given quantity of heat and is proportional to the thermal conductivity (k) and inversely proportional to the specific heat (c) and density (ρ):

$$\alpha = k/c\rho . \tag{4}$$

Thermal diffusivities of some common rocks and soils are given in table 1.

TABLE 1 — THERMAL DIFFUSIVITIES OF SOME COMMON ROCK AND SOIL IN cgs UNITS*

Soils and unconsolidated material:		
	Calcareous earth, 43% water	0.0019
	Quartz sand, medium, dry	0.0020
	Quartz sand, 8.3% moisture	0.0033
	Sandy clay, 15% moisture	0.0037
	Soil, very dry	0.0020–0.0030
	Some wet soils	0.0040–0.0100
	Wet mud	0.0022
	Soil, Lexington, Ky.	0.0021
	Soil, Lexington, Ky. (avg. 0-10' in place)	0.0072
	Gravel	0.0057–0.0062
Rocks:		
	Shale	0.0040
	Dolomite	0.0080
	Limestone	0.0050–0.0110 (0.0080 avg.)
	Sandstone	0.0113–0.0140
	Granite	0.0060–0.0130
Water:		
	At 0° C	0.00131
	At 8° C	0.00169
	Commonly used average	0.00143
Air:		
	At 0° C and 1 atm.	178,000.0000

*From Ingersoll, Zobel, and Ingersoll, 1954; Lovering and Goode, 1963; Penrod, Elliott, and Brown, 1960; National Research Council, 1927.

The mathematical treatment of temperature fluctuations at various depths in the earth is based on the assumption that surface temperature changes generate a heat wave that can be approximated by a sine curve. The distribution of temperature in the earth will approximate a sinusoidal curve and the temperature range at any given depth can be calculated from the following equation, assuming heat moves into the earth material, a semi-infinite body, only by conduction (Ingersoll, Zobel, and Ingersoll, 1954; Lovering and Goode, 1963):

$$T_x = 2T_r e^{-x\sqrt{\pi/\alpha P}} \tag{5}$$

where T_x is the temperature range at depth x, T_r is half the total temperature range at the surface, P is period, and α is diffusivity. The above equation is in cgs units: x in centimeters and P in seconds.

Table 2 (after Lovering and Goode, 1963) is calculated from equation (5). It shows the effective depth of temperature variation due to diurnal and annual atmospheric temperature variation for various values of diffusivity. Temperature variations are given as a percentage of temperature fluctuations of the surface. In central Illinois, the mean monthly temperature fluctuation is about 50° F (28.0° C); 0.1 percent of this is 0.05° F (0.03° C). For most considerations, this may be the effective limit of the surface temperature fluctuation.

TABLE 2 — DEPTH AT WHICH ANNUAL AND DIURNAL TEMPERATURE FLUCTUATIONS ARE 0.1 PERCENT OF SURFACE FLUCTUATIONS, FOR VARIOUS VALUES OF DIFFUSIVITY*

Diffusivity	Depth to nearest 0.1 foot at which diurnal range is 0.1% of surface range	Depth to nearest foot at which annual range is 0.1% of surface range
0.0016		29
0.0025	1.9	36
0.0036	2.0	43
0.0049	2.6	50
0.0064	3.0	58
0.0081	3.4	65
0.0100	3.8	72
0.0121	4.2	79
0.0144		86
0.0169		94
0.0196		101
0.0225	5.7	108

*After Lovering and Goode, 1963.

The effect of nonperiodic surface temperature fluctuations also can be approximated using equation (5). This is done by assuming that the long-duration warm or cold period is one-half cycle of a periodic wave. Thus, a one-week heat wave, during which the average temperature is 10° C (18° F) above the monthly average, has values of T_r = 10° C (18° F) ~~x 2~~ and P = (2 x 7 x 86,400). Table 3 shows the approximate temperature or fluctuation, at various depths, in rocks of different diffusivities for a one-week heat wave.

TABLE 3 — APPROXIMATE TEMPERATURE RANGE IN DEGREES FAHRENHEIT (° F) AND CENTIGRADE (° C) CAUSED BY A ONE-WEEK HEAT WAVE OF 10° C (18° F) ABOVE MONTHLY AVERAGE*

	Depth (in feet)				
Diffusivity	3.3	6.6	9.9	13.2	16.5
0.0049	2.16° C (3.9° F)	0.23° C (0.4° F)	0.02° C (0.04° F)		
0.0100		0.89° C (1.6° F)	0.19° C (0.3° F)	0.04° C (0.07° F)	
0.0144		1.65° C (3.0° F)	0.45° C (0.8° F)	0.13° C (0.2° F)	0.04° C (0.07° F)

*After Lovering and Goode, 1963.

The temperature resulting from the annual wave front can be calculated for any given point beneath the land surface. To derive such an equation, it is necessary to assume that the soil is homogeneous, its surface flat, and the flow of heat in a direction perpendicular to the surface (Ingersoll, Zobel, and Ingersoll, 1954; Lovering and Goode, 1963; Penrod, Walton, and Terrell, 1958; Penrod, Elliott, and Brown, 1960). It is necessary to solve the Fourier heat equation:

$$\frac{\delta T}{\delta t} = \alpha \frac{\delta^2 T}{\delta x^2}, \tag{6}$$

subject to the boundary condition:

$$T = T_r \sin \frac{2\pi}{P} t, \text{ at } x = 0 \tag{7}$$

where T is temperature, T_r is surface temperature range, t is time, P is period, α is thermal diffusivity, and x is the distance from the surface. A particular solution to equation (6) is given by the equation:

$$T_x = T_m + T_r e^{-x\sqrt{\pi/\alpha P}} \left[\sin\left(\frac{2\pi}{P} t - x\sqrt{\pi/\alpha P} \right) \right] \tag{8}$$

where T_x is the temperature at depth x and T_m is the mean annual temperature.

Observed soil temperatures closely follow the theoretical curve resulting from equation (8). Penrod, Elliott, and Brown (1960) and Flucker (1958) made a series of soil temperature measurements over a 5-year period and evaluated equation (8) empirically. The main problem involved is the measurement of T_m, which is the surface soil temperature and is slightly different from mean annual air temperature. Penrod, Elliott, and Brown (1960) found at Lexington, Kentucky, an average soil temperature of 0.67° C (1.2° F) above mean annual air temperature; Lovering and Goode (1963) in the East Tintic District, Utah, found differences as great as 7.78° C (13.9° F) between surface soil temperatures and air temperatures. After smoothing the air temperature curves, they obtained an average soil temperature of 1.90° C (3.4° F) above mean annual air temperature. Flucker (1958) measured soil temperature to a depth of 10 feet over a period of 5 years at College Station, Texas, and found the average soil temperature to be 2.91° C (5.2° F) above the average air temperature; of significance was the fact that soil temperatures decreased continuously with depth, and the soil temperatures were higher than the average soil temperature at a depth of 2 feet or less, while at 3 feet and more they were below average soil temperature.

Many of the differences and discrepancies between soil temperature measurements from area to area can be attributed to the same factors that cause the differences between air and soil temperatures. These factors are soil cover and color, prevalence of sunshine, wind, snow cover (an insulation from further temperature effects), and soil moisture. Frost in the soil can have a significant effect, as it will hold soil temperature near 0° C (32° F) during cold periods and cause a significant lag in the spring temperature rise because of the latent heat of fusion of the ice.

Because they were measured at too shallow a depth, soil temperatures at Champaign-Urbana (table 4) cannot be fitted with confidence to theoretical data. However, the Champaign-Urbana data suggest that equations (5) and (8) are of the correct form (fig. 2) for these data.

TABLE 4 — MEAN MONTHLY SOIL AND AIR TEMPERATURES AT CHAMPAIGN-URBANA TO THE NEAREST DEGREE FAHRENHEIT*

		Depth below surface			
Month	Air	4"	12"	24"	36"
January	27	27	33	38	41
February	29	27	32	37	38
March	40	33	38	39	41
April	51	44	47	47	48
May	62	57	52	56	54
June	71	68	68	66	61
July	76	72	71	70	67
August	73	72	71	71	69
September	67	65	64	68	67
October	55	44	53	61	60
November	42	38	47	51	52
December	30	30	37	43	44
Average	52.0	48.1	51.1	53.9	53.5

*After Changnon, 1959.

Comparing these data with that of Penrod, Elliott, and Brown (1960) and Flucker (1958), the probable mean annual surface soil temperature is about

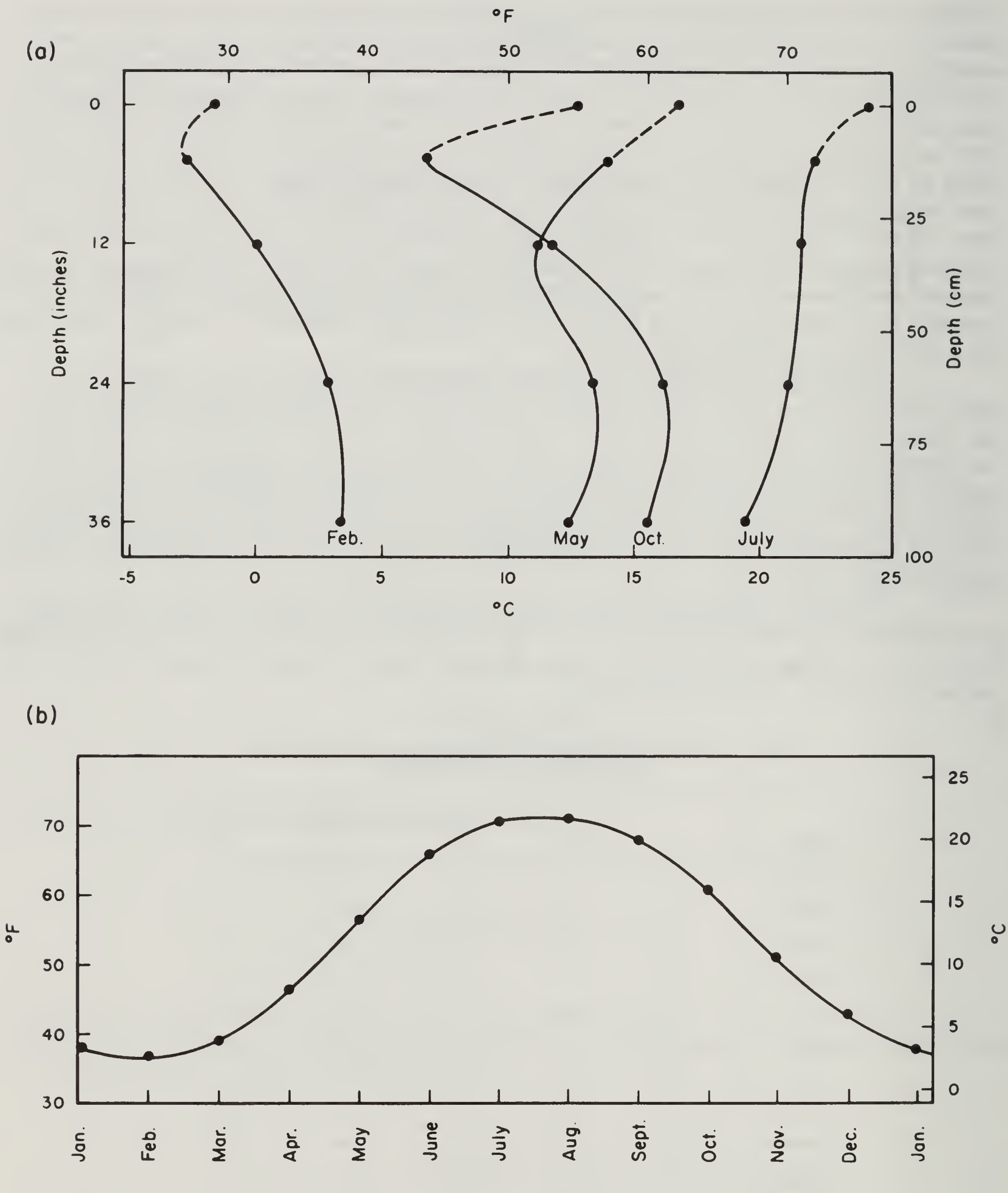

Figure 2 - (a) Some temperature depth profiles at Champaign-Urbana (after Changnon, 1959).
(b) Mean monthly temperature 24 inches below surface at Champaign-Urbana (after Changnon, 1959).

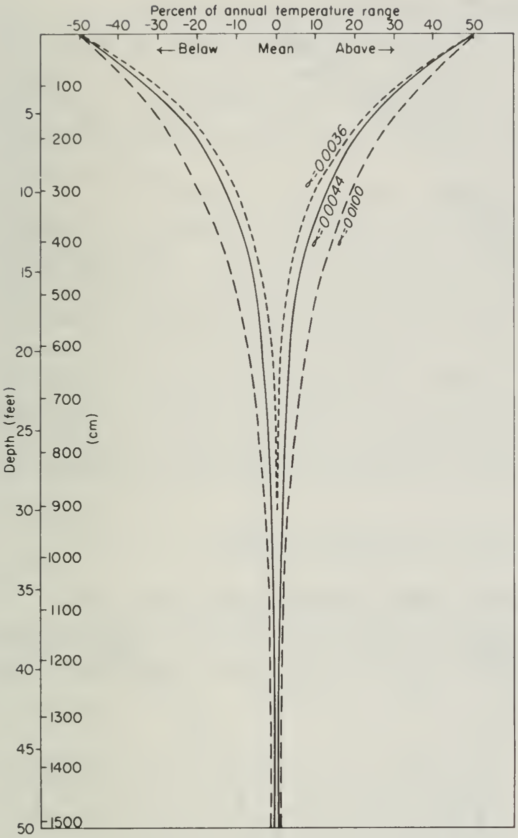

Figure 3 - Maximum temperature fluctuation at depth in three soils of different thermal diffusivity (α). Temperature fluctuation is given as percentage of surface annual temperature range (after Lovering and Goode, 1963).

53.7° F (12.15° C), 1.7° F (0.95° C) above the mean air temperature. The lower temperatures of the 4- and 12-inch measurements may be due to cooling by evaporation of moisture and the cooling effect of rain, which mostly affects the top of the soil (Flucker, 1958).

From equation (8) a set of curves can be calculated, showing soil temperatures at any depth and any time knowing the surface temperature fluctuation (wave). An envelope that contains all the curves can be calculated, using equation (5). Figure 3 shows three envelopes for different diffusivity values.

Soil temperature measurements may be used to calculate the value of diffusivity (α) of soils in place. Lovering and Goode (1963) present a graph from which values of (α) may be obtained. Using this graph and the soil temperatures at Champaign-Urbana reported by Changnon (1959), the values of diffusivity range from 0.0040 to 0.0080, with 0.0050 the average value. The temperature data are not at great enough depth for accurate diffusivity determinations. Figure 4 shows the envelope and some oı the temperature curves for a diffusivity of 0.0049.

In calculating the soil temperature curves it is assumed that the land surface (annual wave) is the only heat source or sink (that is, all heat is either gained or lost at soil-air interface), which actually is not the case. The rock below is a constant heat source, as is shown by the geothermal gradient. Therefore, the envelopes should be modified to take account of this factor; this can be done graphically. Figure 4 can be modified to meet the geothermal gradient expected in a glacial till in east-central Illinois. Lovering and Goode (1963) made a temperature measurement through the zone affected by the surface wave. Those data show a slight slope of the curves, as might be expected.

TEMPERATURE PROSPECTING

The theoretical equations cited above adequately describe the thermal properties and expected temperatures in a glacial till sequence. The presence of a saturated sand aquifer in a normal sequence of till will upset the existing thermal balance, or, rather, create a new thermal system, which is different from areas where no shallow aquifers exist. The low values of (α) for water (0.0014) and quartz

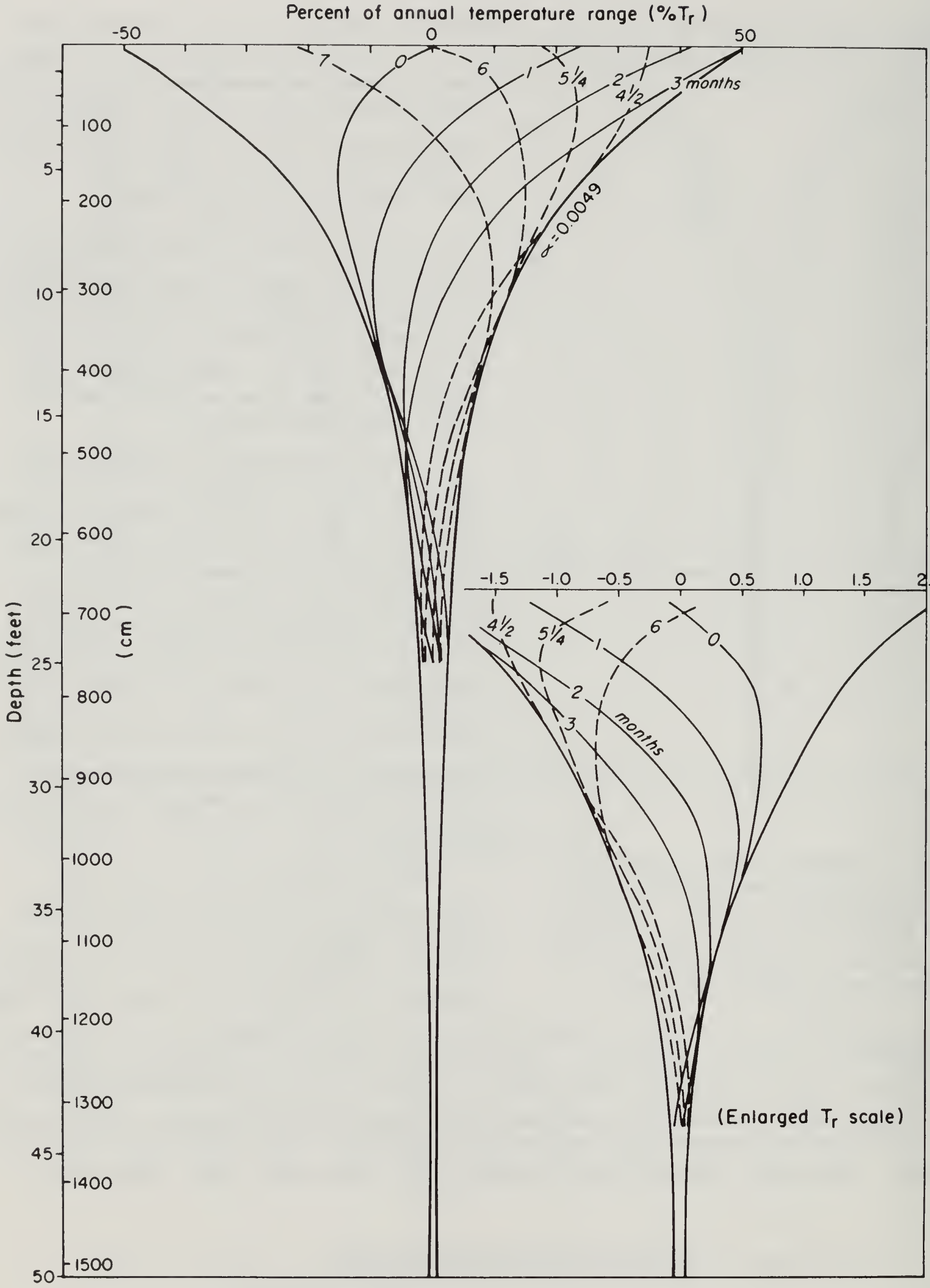

Figure 4 - Theoretical depth-temperature curves and maximum temperature fluctuation for soil with thermal diffusivity (α) of 0.0049. The curves show the soil temperature at various times (in months) after the spring crossover when soil temperature at the surface is equal to the annual mean temperature. Temperatures are given as percentage of surface annual temperature range (after Lovering and Goode, 1963).

sand (0.0020 to 0.0033) and the high specific heat of water make the aquifer a heat sink (or source at certain times). Lovering and Goode (1963), in attempting to detect high thermal gradients from ore bodies in the East Tintic District, Utah, found that a perched water layer (water table was at 1000 feet) would absorb the heat generating in oxidizing ore bodies and therefore reduce the gradient detected at the surface.

The fact that near-surface soil temperatures will be affected by the change in material is obvious; however, the size and nature of the anomaly is of some question. The anomaly either must be large enough to be distinguishable from extraneous temperature fluctuations, or the extraneous fluctuations must be evaluated and/or eliminated in some way from the field data. Much of this can be accomplished by taking readings at a single instant in time or by using a base station to record changes in soil temperatures with time. In all cases, the aquifer will act as a flexure point in the temperature gradient curve, serving as the maximum depth of penetration of surface temperature fluctuation (wave) and the point at which the geothermal gradient is near, or at, mean annual temperature.

Thermal Anomaly Caused by Shallow Aquifers

To evaluate the effect of a shallow aquifer on the soil temperature near the land surface, the following assumptions are made: (1) the buried aquifer is of sufficient extent that the flow of heat between the top of the aquifer and the land surface can be considered as a one-dimensional transfer process at the point where the temperature is measured; (2) the aquifer is overlain by nonwater-yielding material, which can be considered as a slab of uniform thickness and thermal properties and through which heat is transferred between the land surface and the aquifer; (3) the temperature at the upper boundary of the slab (land surface) varies with the sinusoidal yearly temperature wave, and the lower boundary (the aquifer) may be considered a constant temperature boundary.

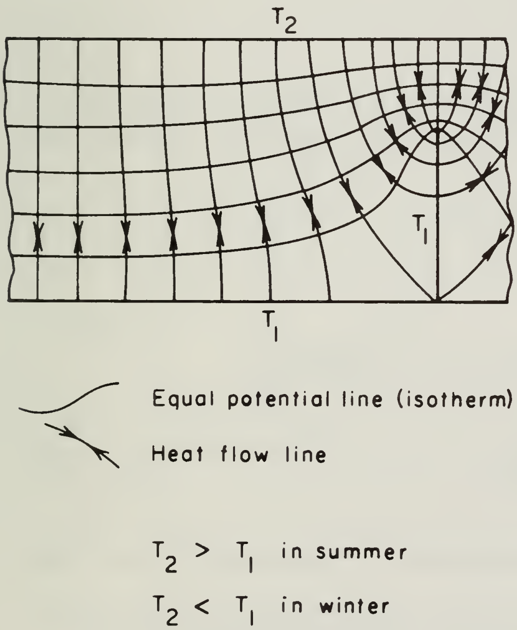

Figure 5 - Isotherms and flow lines for steady heat conduction through a soil (after Ingersoll et al., 1954, p. 203).

The effect of a shallow aquifer on the temperature of the overlying material can be illustrated graphically by a series of theoretical isotherm and flow lines. The illustration in figure 5, from Ingersoll, Zobel, and Ingersoll (1954, p. 203), graphically shows the heat flow through a wall as affected by the presence of an internal projecting rib. It is assumed that the rib has a high conductivity as compared with the wall so that it is an isothermal surface taking the temperature T_1 of the surface of the wall that it adjoins. This would be similar to an aquifer in the zone affected by surface temperature fluctuations where during the cold and hot months the aquifer maintains a temperature that is the same as the temperature at some greater depth.

Quantitative analysis of the expected anomaly can be approached by using the general heat equation for a permanent heat source (Ingersoll, Zobel, and Ingersoll, 1954):

$$T = \frac{S' \, x^{(2-n)}}{2\pi^{n/2}\alpha} \int_{\frac{x}{2\sqrt{t\alpha}}}^{\infty} \beta^{(n-3)} e^{-\beta^2} \, d\beta \tag{9}$$

where T is the temperature at depth x at time t, S' is the strength of the heat source, α is diffusivity, and β is a variable of integration.

For a one-dimensional flow—that is, in the vertical direction only between the heat source and land surface—n = 1 and equation (9) becomes

$$T = \frac{S' \, x}{2\alpha\pi^{\frac{1}{2}}} \int_{\frac{x}{2\sqrt{t\alpha}}}^{\infty} \beta^{-2} e^{-\beta^2} \, d\beta \, . \tag{10}$$

The strength of the heat source, S', is proportional to the quantity of heat, Q', lost through the surface and inversely proportional to the density and specific heat (c) of the material transmitting the heat:

$$S' = \frac{Q'}{c\rho} \, . \tag{11}$$

If the material is uniform, the quantity of heat, Q', is obtained by

$$Q' = \frac{\Delta T t}{x/kA} \tag{12}$$

where x is the interval of measurement of the temperature difference (ΔT), and k is the thermal conductivity.

Equation (10) describes the unsteady flow of heat through an infinite slab that is uniformly and suddenly heated at one surface while the other surface is held at a uniform temperature. The equation shows that there is no steady state because the value of the integral becomes very large as time approaches infinity.

The steady-state equation can be derived from the basic Fourier equation, which for one dimension is

$$\frac{d^2T}{dx^2} = 0 \, . \tag{13}$$

When integrated, the equation becomes

$$T = Bx + C \, . \tag{14}$$

If the boundary conditions $T = T_1$ at $x = \ell$ and $T = T_2$ at $x = m$ are inserted in equation (14) where m and ℓ are, respectively, the distances from the reference datum (yz plane) of the upper and lower surfaces of the slab surfaces (soil-air interface and aquifer-overburden interface), equation (14) becomes

$$T = \frac{mT_1 - \ell T_2}{m - \ell} - \frac{(T_1 - T_2)r}{m - \ell} \tag{15}$$

where T is the temperature at any point r within the slab (fig. 6).

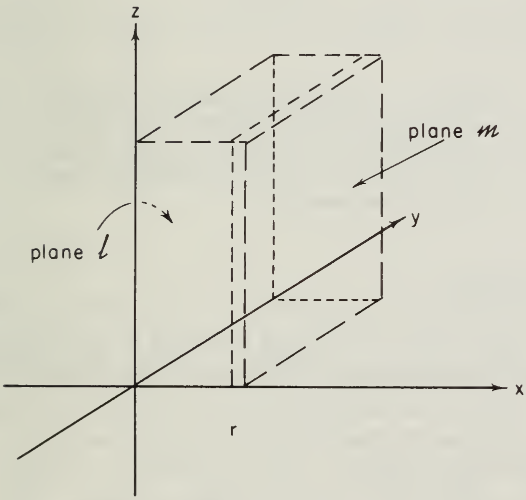

Figure 6 - Geometry of the slab used in equation 15. Plane m is the air-soil interface, plane ℓ is the aquifer-overburden interface, and plane r is a random plane in the slab parallel to plane m where the temperature is calculated.

The following assumed reasonable values for the parameters of equations (5) and (15) are based on data obtained by McGinnis, Kempton, and Heigold (1963), Loofbourow (1966), Ingersoll, Zobel, and Ingersoll (1954), and Clark (1966):

diffusivity (α), 0.005 cgs units
depth to top of the aquifer (x), 500 cm
density of overburden (ρ), 2.35
specific heat of overburden (c), 0.2
temperature difference (ΔT), 15° C
thermal conductivity of overburden (k), 0.002.

When these values are substituted in equation (15), a temperature increase of 1.5° C is obtained. This is the maximum temperature anomaly that can be expected during the coldest and warmest months of the year. Figure 7 shows the monthly average temperature variations at 50 cm below the land surface from an aquifer buried 500 cm below the land surface, assuming steady-state conditions for each month for the average monthly deviation from the mean annual temperature, based on temperatures taken at Champaign, Illinois. If equation (10) is set equal to equation (15), the approximate time required before a steady-state condition is reached can be solved. For the conditions previously described—that is, for x = 500 cm—a time of 1.47 x 10^6 seconds (about 17 days) is required to reach equilibrium. For a larger x (greater depth) the time is greater, and, conversely, for a smaller x the time is less.

In addition, the aquifer should act as a heat sink (Lovering and Goode, 1963), interrupting the flow of geothermal heat to the earth's surface. The resulting decrease of the geothermal gradient should cause a slight cooling of the surface soil, and this is added to the previously described effect of the heat transfer. If a temperature gradient in the glacial till of 54.69° C per kilometer (3° F per 100 feet) is used, a decrease in soil temperature of about 0.05° C (0.09° F) is estimated. This increases the summer anomaly and decreases the winter anomaly, and possibly could be used to detect deeper aquifers.

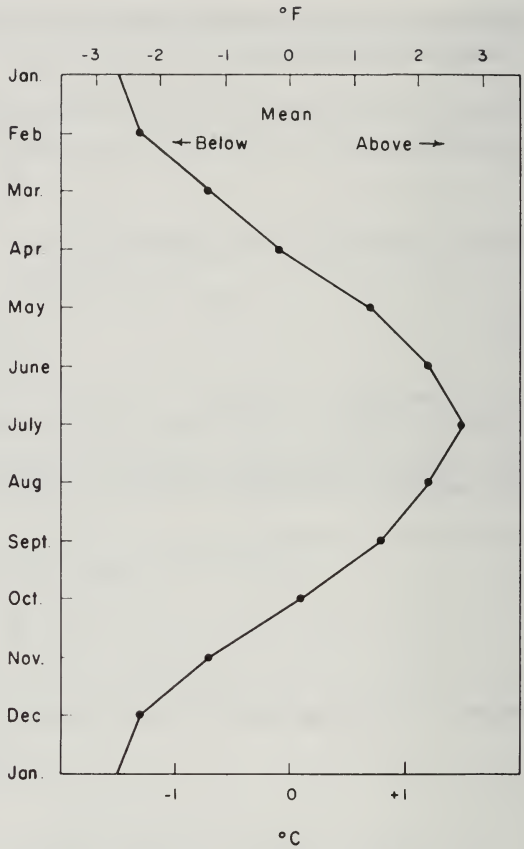

Figure 7 - Calculated departure from normal temperature of soil at 50 cm depth, resulting from an aquifer at depth of 500 cm and 15° C maximum difference from surface temperature.

Effect of Depth

From inspection of equations (9), (10), and (15), it can be seen that increasing the depth to the aquifer will have the effect of decreasing the surface temperature anomaly. The curve in figure 8 was made by solving equations (10) and (15) for numerous slab thicknesses (with the IBM 7094 computer at the University of Illinois) for the temperature at 50 cm (19.7 inches) below the surface, which is held at uniform temperature in a material with the properties assumed in the previous section. The curve shows the relation of depth to the size of the anomaly that can be expected at 50 cm below the land surface. To these values must be added the loss of heat due to interruption of geothermal heat flow by the aquifer. The maximum effective limit of the depth of an aquifer that can be detected is probably between 1000 and 2000 cm (32.8 to 65.6 feet), depending on the ability of the operator to distinguish the desired anomaly from surface temperature changes caused by such surface factors as changes in soil color and character, vegetation, and direction and amount of slope.

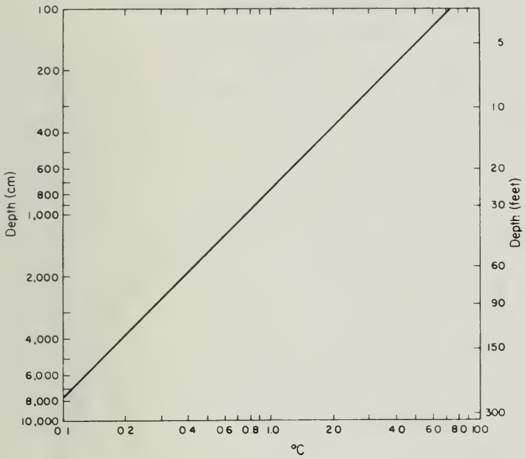

Figure 8 - Relation of depth of an aquifer with a temperature 15° C different from that at surface to the maximum temperature anomaly theoretically observed at a depth of 50 cm.

Depth Estimates

From the discussion of the effect of depth of the buried body on the surface temperature and figure 8, it is immediately obvious that the size of the anomaly is directly related to depth of burial. Actual field observation confirms this general relationship. However, the data also suggest that thickness and permeability of the aquifer may affect the size of the anomaly.

A more accurate method of determining depth can be obtained using equations (8) and (15), although this is subject to similar problems as is the use of the size of the anomaly to determine depth; that is, it ignores the aquifer properties. This method involves the difference in temperature obtained by two simultaneous readings at different depths at the same site. The temperature difference will be equal to the temperature difference due to the normal surface effect (equation 8) plus the temperature difference due to heat flow from the buried aquifer (equation 15):

$$\Delta T = \left\{T_m + T_r\left[e^{-x_1\sqrt{\pi/\alpha P}}\sin\left(\frac{2\pi t}{P} - x_1\sqrt{\pi/\alpha P}\right)\right] + \left[\frac{mT_1 - \ell T_2}{m - \ell} - \frac{(T_1 - T_2)x_1}{m - \ell}\right]\right\}$$
$$- \left\{T_m + T_r\left[e^{-x_2\sqrt{\pi/\alpha P}}\sin\left(\frac{2\pi t}{P} - x_2\sqrt{\pi/\alpha P}\right)\right] + \left[\frac{mT_1 - \ell T_2}{m - \ell} - \frac{(T_1 - T_2)x_2}{m - \ell}\right]\right\} \tag{16}$$

where x_1 and x_2 are the depths of the temperature readings, and ΔT is the temperature difference between the readings at x_1 and x_2. Rearranging and solving for the bottom of the slab m, yields

$$m = \ell + \frac{(x_1 - x_2)\quad(T_1 - T_2)}{T_r\left\{\left[e^{-x_1\sqrt{\pi/\alpha P}}\sin\left(\frac{2\pi t}{P} - x_1\sqrt{\pi/\alpha P}\right)\right] - \left[e^{-x_2\sqrt{\pi/\alpha P}}\sin\left(\frac{2\pi t}{P} - x_2\sqrt{\pi/\alpha P}\right)\right]\right\} - \Delta T} . \tag{17}$$

Equation (16) can be solved for various thickness and temperature differences between the surface and the aquifer assuming reasonable values for the factors involved. Figure 9 shows a plot of the increased temperature resulting from a buried body with a temperature 15° C different from the surface, and also the expected difference between readings at 50 and 100 cm, taking into account heat from the anomalous body and normal temperature increases (equation 8).

Several factors are difficult to evaluate because they cannot be directly measured. Temperature difference between the surface and aquifer (T_1-T_2) is the most critical factor and one that is not directly measurable. Values of diffusivity (α) and time of the year after the temperature crossover (t) can be reasonably estimated, and small differences are not critical in the final results.

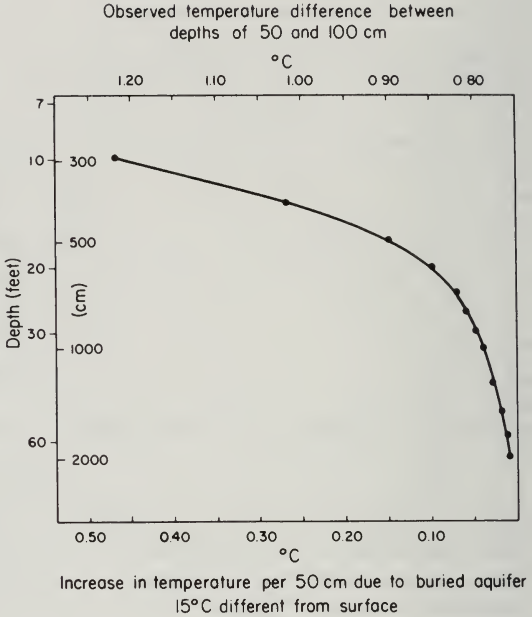

Figure 9 - Temperature increase per 50 cm (equation 8) and observed temperature difference between depths of 50 and 100 cm (equation 16) resulting from an aquifer with a temperature 15° C different from that at the surface and occurring at various depths. The curve can be used to estimate the depth of the aquifer.

PREVIOUS WORK ON TEMPERATURE PROSPECTING

A number of studies have been made of geothermal gradients. Work on the use of temperature prospecting has been limited. Van Orstrand (1940) measured temperatures in wells in the Salt Creek Field in Wyoming and found an increase in geothermal gradient over the structure; he attributed the increased gradient to the upwarping of hotter strata toward the land surface. Guyod (1946) also noted similar anomalies associated with salt domes and suggested that thermal measurement might be used to detect salt domes; Dobrin (1952) suggested that the reason for this is the very high thermal conductivity of the salt. Stallman (1965) and Bredehoeft and Papadopulos (1965) used soil temperature changes and thermal profiles to determine vertical velocity of ground-water flow in the soil and the vertical permeability of the soil.

Kintzinger (1956) noted an anomaly of 12° C (21.6° F) over a hot water area near Lordsburg, New Mexico, where no surface expression of the hot water was present, but super-heated water was encountered in wells at a depth of 78 feet. Lovering and Goode (1963) studied the possibility of detecting oxidizing ore bodies from abnormally high geothermal gradients in the East Tintic District, Utah, and concluded that it was possible, but not practical. Strangway and Holmer (1966), using infrared photography and soil temperature surveys, described thermal anomalies over several geologic structures and thermal water areas; a 10° F (5.6° C) anomaly was found over an area where 150° F (~~84.0°~~ C) water was reported at a depth of 450 feet (137 meters).

65.6°

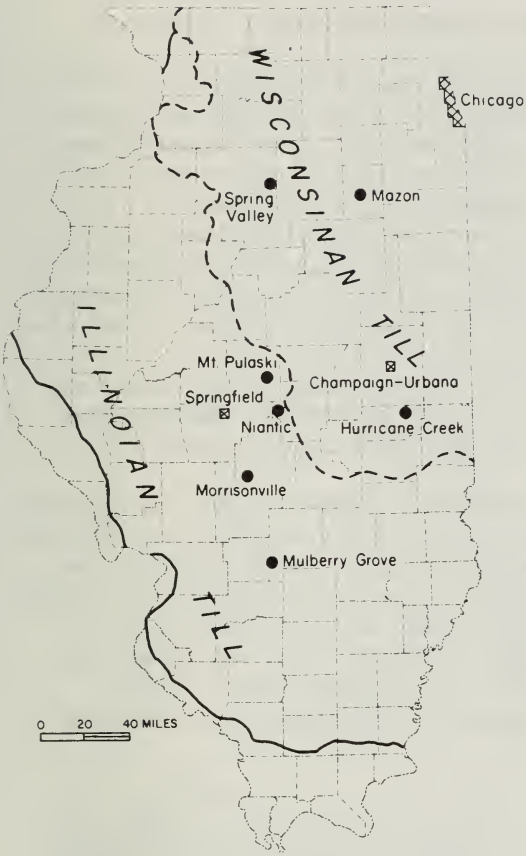

Figure 10 - Location of the field temperature studies.

Birman (1965) describes a potential method of geothermal prospecting in which temperature probes are buried at a depth of about three meters, well below the zone affected by diurnal temperature variations. Readings are made about a month apart. This method measures the effect of the aquifer on the annual temperature wave.

Carr and Blakely (1966) employ a technique similar to Birman's, but use variation in the diurnal temperature wave. In this method of prospecting, the temperature probe is only 30 cm below the surface and measures the diffusivity (α) of the surface material.

This paper complements previous preliminary studies in Illinois (Cartwright, 1966, 1968).

FIELD STUDIES

A number of field studies have been conducted to verify the theoretical calculations given above. To date, nine field studies have been made, of which seven are discussed in this report (fig. 10); the other two studies were exploratory for new well fields, and the data collected thus far are not adequate enough to confirm the results.

The size of the anomalies encountered in the field closely approximate the theoretical estimates made in the previous section of this paper. However, the change from winter to summer type anomalies and back again may be more rapid and earlier in the seasons than predicted.

Measurement and Instrumentation

For useful temperature prospecting, the anomaly must be detectable. As has been shown, the main source of rapid temperature fluctuation is the diurnal fluctuation, which is effective to a depth of about 3 feet. Strangway and Holmer (1966) made their temperature readings at a depth of 75 to 80 cm (30 to 32 inches) in the soil. In general, diurnal temperature fluctuations are very small at depths of 45 cm (18 inches) or greater, and they probably occur rather slowly. Temperature data for Champaign-Urbana show a summer fluctuation of 5° to 10° F (2.8° to 5.6° C) in the summer and 1° to 4° F (0.6° to 2.2° C) in the winter at a 4-inch depth (Changnon, 1959). At a 12-inch depth, however, the summer fluctuations are about 2° to 3° F

(1.2° to 1.8° C), and winter fluctuations are generally less than 1° F (0.6° C), commonly less than 0.5° F (0.3° C).

Thus, it is possible to eliminate most of the diurnal fluctuations by taking readings at 45 cm (18 inches) or greater in depth. The small temperature changes can be eliminated by taking a large number of measurements at an instant in time. This is not difficult, as the slow fluctuation of temperature at these depths will allow a considerable time lapse, without a significant temperature drift.

If, however, the survey is to be large, made over a considerable span of time, it is advisable to set up a base station at which to measure temperature drift. A second method of controlling temperature drift is to make each series of readings overlap. Longer period fluctuations such as hot or cold spells also can be adjusted in the same manner.

Other factors affecting soil temperatures, such as color of the soil, vegetation, etc., cannot be evaluated systematically, nor quantitative relationships given at the present state of knowledge. These factors can be minimized in the field by the choice of station sites.

The instrument for measuring soil temperatures is quite simple. It consists of a thermister at the end of an aluminum-tipped stake or probe, which can be driven into the ground (fig. 11), and a transistor-amplified bridge circuit (modified from Radio and Electronics, 1963). This is calibrated in the laboratory to convert microvolts to temperature. In order to obtain a whole profile at an instant in time, a series of stakes can be connected to a multistrand wire and read individually by turning a selector switch. In general, however, the soil temperature remains sufficiently constant for a period of several weeks so that readings can be made over this span of time with a single probe.

Hurricane Creek

Hurricane Creek is an alluviated valley 8 miles southeast of Charleston, Illinois (fig. 10). The valley is cut into the Illinoian till plain and trends south from the outermost Wisconsinan Moraine, the Shelbyville. The stream, a tributary to the Embarras River, carried outwash gravel, sand, and silt away from the Shelbyville Moraine. The area investigated lies about 1½ miles south of the moraine.

The alluvial fill of the valley is about 45 feet thick (13.7 meters). The maximum thickness of the aquifer is about 38 feet (11.6 meters). The deposit is underlain by impermeable Pennsylvanian age shale and sandstone and overlain by silty, sandy alluvium. The aquifer was located and outlined by an electrical earth resistivity survey (Buhle, 1953). Several large-capacity water wells have been developed in the aquifer by the Forest Oil Company.

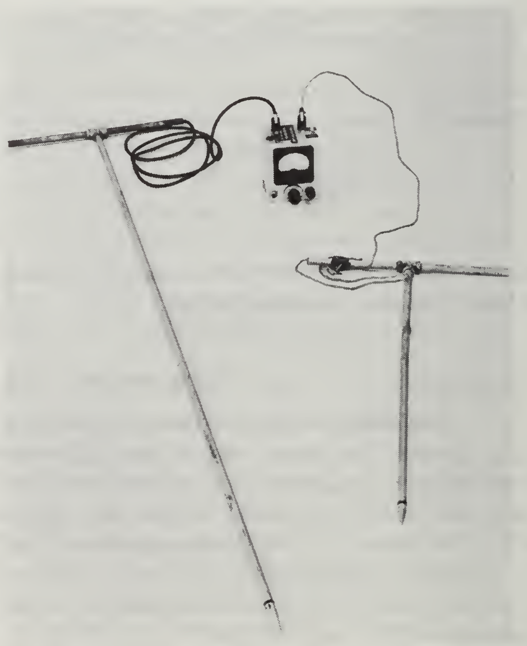

Figure 11 - Instrument used in the field temperature studies (plans modified from Radio and Electronics, 1963).

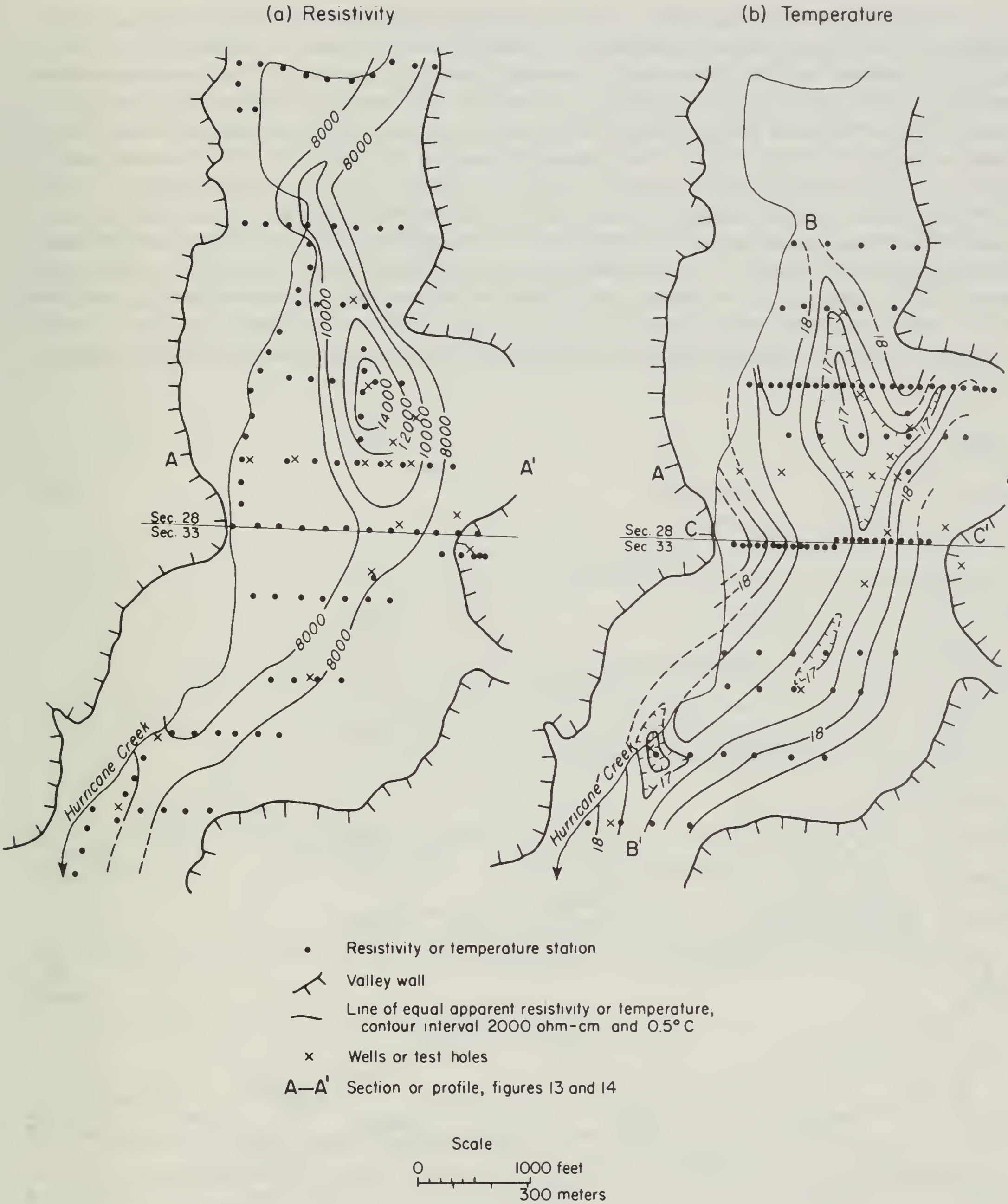

Figure 12 - (a) Resistivity (apparent) at 50-foot electrode spacing (from Buhle, 1953), and
(b) Temperature 18 inches below land surface in June 1966 on Hurricane Creek flat, sec. 28 and 33, T. 11 N., R. 10 E., Cumberland County, Illinois.

The resistivity map (fig. 12a) is an isoresistivity map using the apparent resistivity at an electrode spacing of 50 feet in the Wenner configuration. The main part of the aquifer lies within the 8000 ohm-cm contour. The isotemperature map (fig. 12b) of the same deposit is based on 169 readings made during early June 1966. The data are corrected for soil temperature drift to June 8. The contour interval is 0.5° C (0.9° F). The maximum anomaly is about 1.25° C, which is close to values predicted theoretically (the maximum winter anomaly is 0.75° C—1.4° F). The cool anomaly closely fits the resistivity anomaly (fig. 12a, b) and test boring data relative to the location of the aquifer. The two large production wells presently in use lie within the two strong temperature anomalies. The small temperature low on the western side of the valley coincides with a small sand body.

The geologic cross section at A-A' (fig. 13) is made from test boring and resistivity data. Directly above the geologic cross section are two temperature

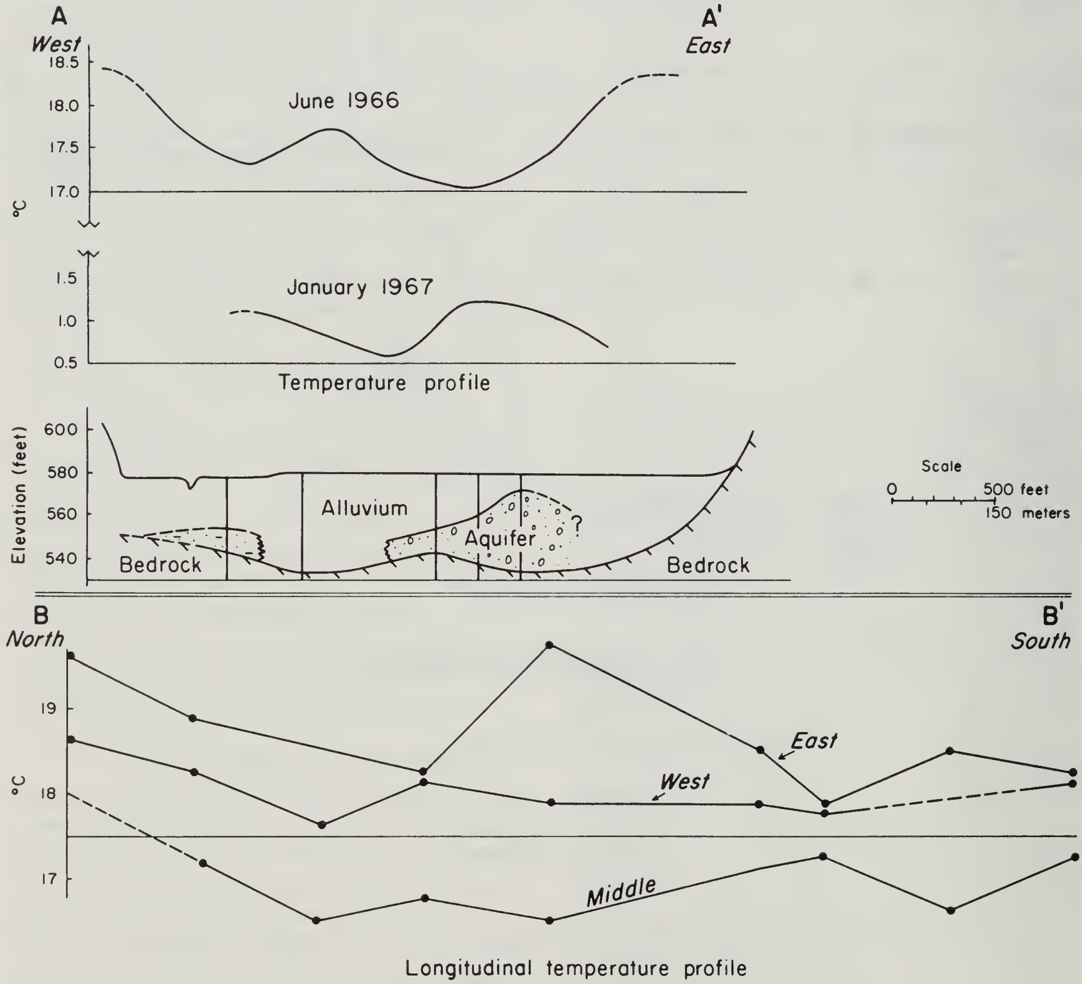

Figure 13 - Temperature profiles (winter and summer) and geologic cross section along A-A' (fig. 12) and north-south longitudinal temperature profiles along B-B' (fig. 12), Hurricane Creek.

profiles, one made in the summer and one in the winter. The temperature data are nearly an exact fit. A series of longitudinal temperature profiles were made down the valley (B-B', fig. 13). The temperatures in the center of the anomaly are compared with the temperatures on the eastern and western sides. Not enough is known of the hydrogeology of the deposit to relate the variations in temperature with variations in permeability; however, the limited data available suggest this is a possible explanation.

The temperature profiles at C-C' (fig. 14) are approximately 180 meters (approximately 600 feet) south of A-A'. The three profiles, made about a week apart between May 26 and June 8, show the general rise in temperature of the soil. These profiles show that the results are fairly reproducible. A fourth profile made in the winter is also shown. The May 26, June 8, and winter profiles match quite well considering the differences in the number of stations of the profiles. The June 3 profile is a bit erratic, and the thermal high near the center is almost indis-

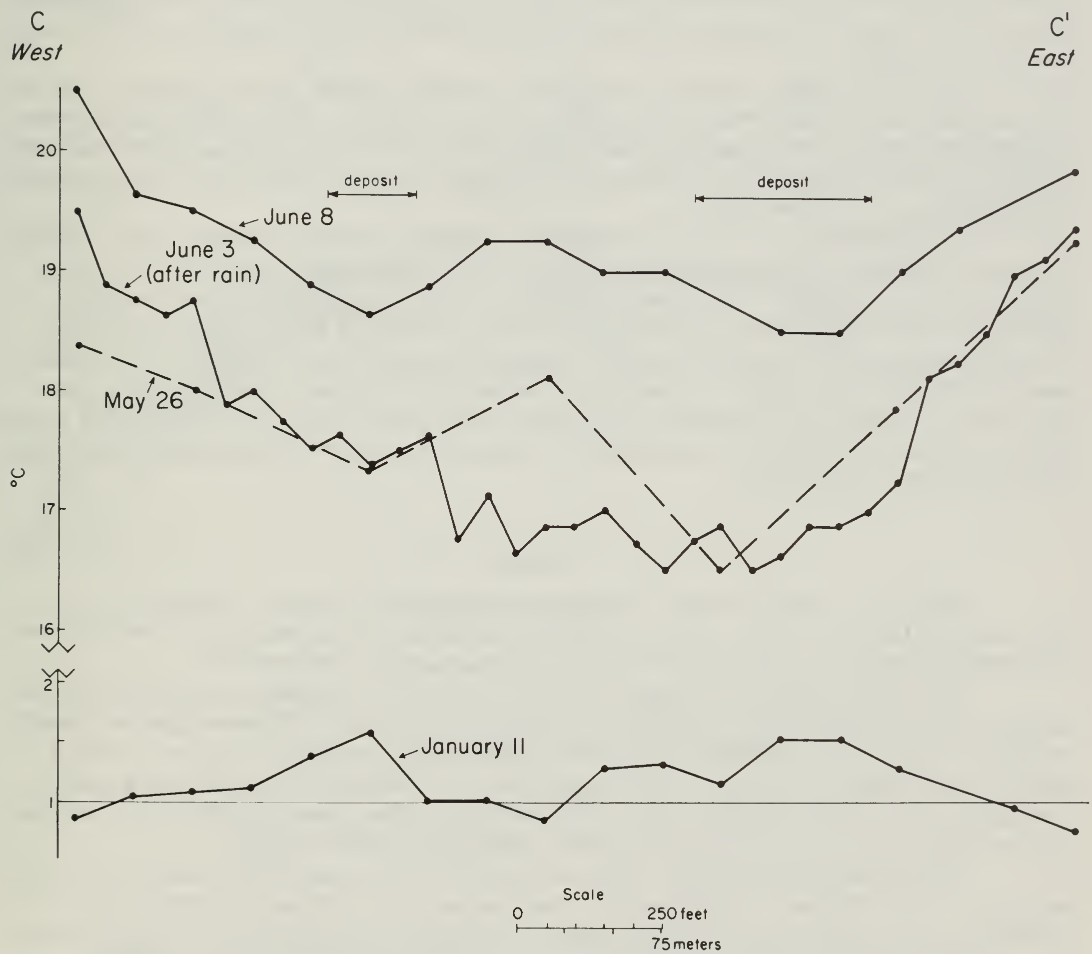

Figure 14 - Temperature profiles (summer and winter) across the valley along C-C' (fig. 12), Hurricane Creek.

tinguishable. This is attributed to the fact that during the night there was a moderate rain, which strongly affected the soil temperature, although it did not completely eliminate the anomaly.

Niantic

Niantic is in central Illinois about 25 miles east of Springfield (fig. 10). The site is on an outwash plain in front of the Shelbyville Moraine, about half a mile west of the moraine. In the western part of the area, the gravel deposits are confined to the valley of the small stream; the stream valley is essentially indiscernible in the eastern part of the mapped area. The aquifer lies below 15 to 20 feet (4.5 to 6 meters) of alluvium, and overlies Illinoian glacial till.

The area has been one of extensive resistivity surveys (Buhle, 1953; Emery, 1942); the resistivity and temperature maps of the deposit are shown in figure 15. The two maps are strikingly similar but not identical. Test drilling suggests that neither map precisely delineates the aquifer, but they are close. The cross section at A-A' (fig. 16) shows a close correlation between the geology and the temperature anomaly.

One of the most prominent features of both maps is the increasing size of the anomaly toward the east, the source of the outwash. This is also illustrated in the longitudinal cross section and temperature profile (fig. 17). This suggests a close relation between the size of the anomaly and the coarseness (and probably permeability) of the material.

The anomaly of 1° to 1.5° C in the summer is approximately what is theoretically expected. The winter anomaly of 1° C is also reasonable. The winter curves at A-A' (fig. 16) illustrate a problem of working in winter months when there is considerable frost in the ground. The profile at 45 cm (18 inches) was strongly affected by the frost, which extended almost to that depth. The profile made at 60 cm (24 inches) the same day shows only a small anomaly. The profile made at 100 cm (39 inches) five weeks later, after the construction of a longer probe, shows an anomaly as expected. This points out the need to have the probe well below the frost line.

Mazon

Mazon is a small town 50 miles southwest of Chicago, situated in a Pleistocene lake flat (fig. 10). The area is extremely flat with less than 10 feet (3 meters) of relief, except where creeks have cut into the surface.

The aquifer lies at a depth of about 15 feet (4.5 meters), and there is no surface expression of its presence. The aquifer is fine- to coarse-grained sand, which coarsens to a gravelly sand at the base. The maximum thickness of the aquifer is 13 feet (4 meters). It is underlain by Wisconsinan glacial till and overlain by late Wisconsinan lake silts and clay. The exact origin of the aquifer is not known, but it is presumed to be a pre-lake stream deposit.

The resistivity map (fig. 18a) was made at the time of the discovery of the aquifer in 1938 (Buhle, 1938), with a small amount of additional work in the southern part of the area in 1966, after the mapping of the deposit by temperature methods. The temperature anomaly map (fig. 18b) is based on data from 154 stations taken a number of times, making it difficult to correct the temperatures to any one date.

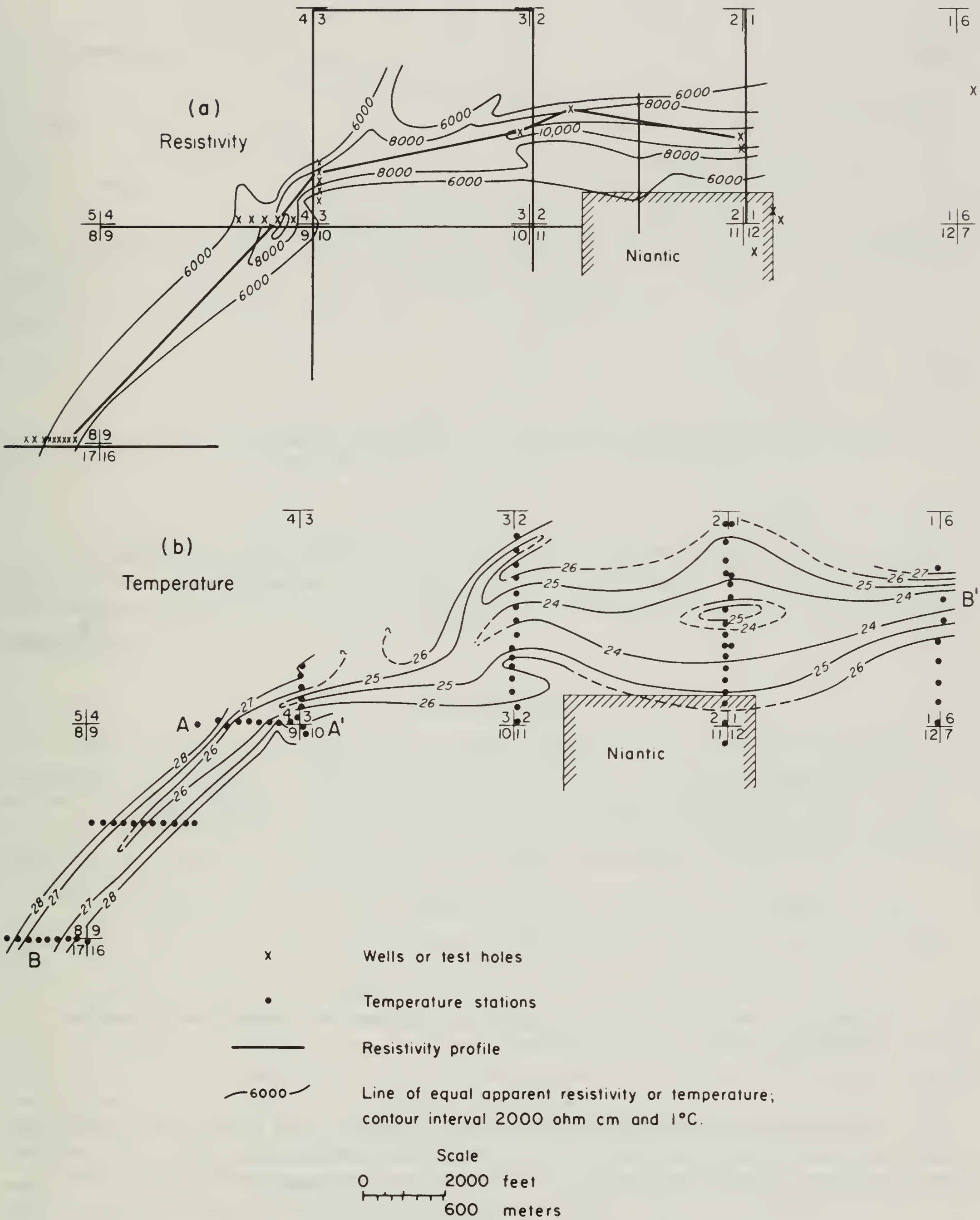

Figure 15 - (a) Resistivity (apparent) at 40-foot electrode spacing (from Buhle, 1953, and Emory, 1942), and
(b) Temperature 18 inches below land surface in the Niantic area, T. 16 N., R. 1 E., Macon County, Illinois.

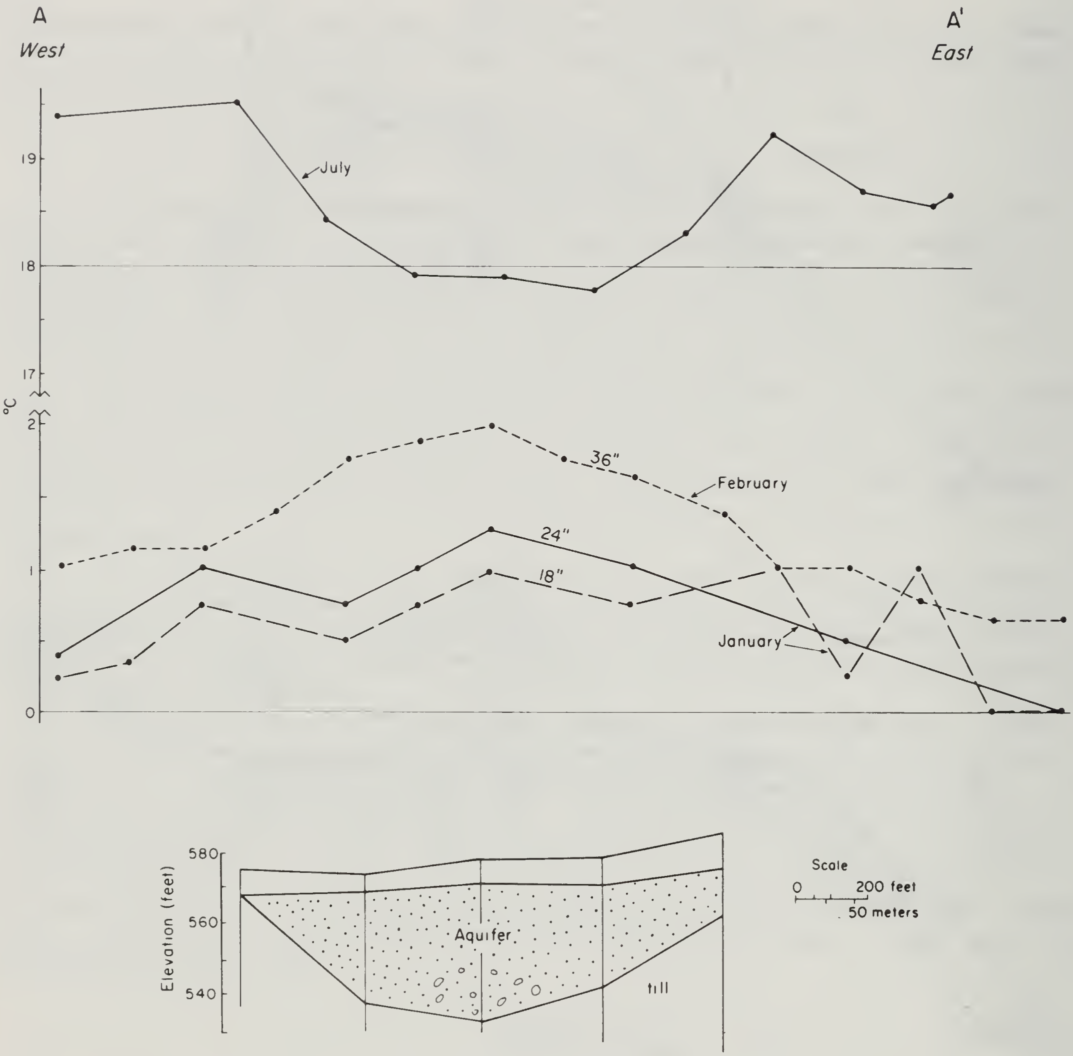

Figure 16 - Temperature profiles (winter and summer) and geologic cross section along A-A' (fig. 15), Niantic area.

A geologic cross section of the deposit at B-B' and three temperature profiles made during different times of the year are shown in figure 19. The profiles show a winter anomaly of about 1.5° C and a summer anomaly of about 1.75° C. The winter profile (3/28/66) shows a normal warm anomaly over the deposit. However, the two summer profiles (5/25/66 and 7/6/66) show a warm spot in the cold anomaly in the immediate vicinity of the pumping wells. This is attributed to the draining of the moderately permeable silt, which overlies the aquifer. Profiles at A-A' and C-C' (fig. 19), 1000 feet (300 meters) north and south of the well field, show normal cool summer anomalies.

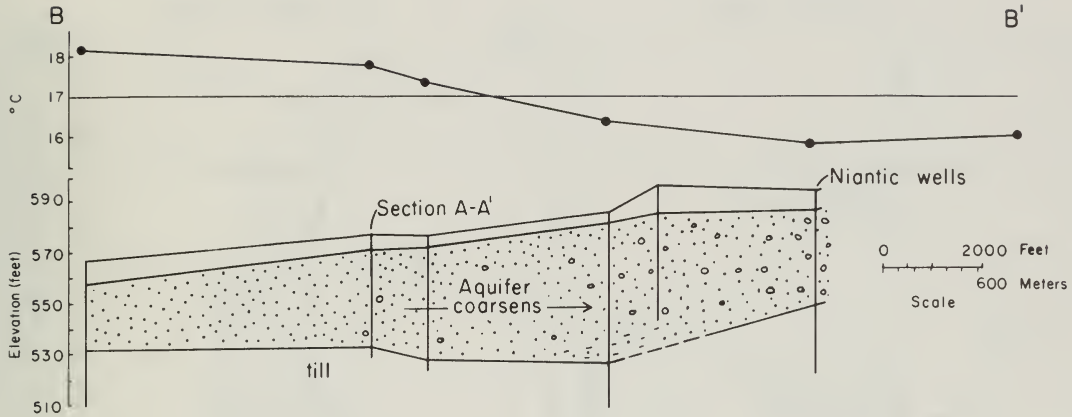

Figure 17 - Longitudinal temperature profile and geologic cross section along B-B' (fig. 15) along the middle of the deposit, Niantic area.

The longitudinal profile, D-D' (fig. 19), shows the difference in temperatures between the center and the edges of the deposit. For the purpose of this illustration, an attempt was made to correct the temperature to the 7/6/66 temperature values.

Morrisonville

Morrisonville is a small town on the Illinoian till plain about 25 miles southeast of Springfield in central Illinois (fig. 10). The aquifer is thought to be an ice-crevasse deposit, which has been traced almost continuously over a distance of 75 miles by electrical earth resistivity methods. The crevasse deposit seems to bear some relationship to present-day topography, apparently acting as a focus for the location of streams.

In the Morrisonville area, the aquifer has a thickness of about 20 feet (6 meters). It is underlain by Illinoian glacial till and overlain by 15 to 30 feet (4.5 to 9 meters) of Illinoian till or, where streams have cut through till, by 10 to 13 feet (3 to 4 meters) of silty alluvium.

The map showing the location of the aquifer (fig. 20a) was made using well data (Cartwright, 1962) and a reinterpretation of the original resistivity data (Buhle, 1942); the apparent resistivity map does not give an accurate picture of the deposit. The temperature map (fig. 20b) gives a close approximation of the deposit. By test drilling in the vicinity of the aquifer, it was found that the material in the center of the deposit where the wells were built is much more permeable than the material nearer the edges. The maximum temperature anomaly, about 1° C, is in the immediate vicinity of the present city well field.

The cross section (fig. 21) shows the relationship of the aquifer, as interpreted from geologic and resistivity data, to the temperature and apparent resistivity profiles. The most permeable part of the deposit again seems to coincide with the greatest temperature anomaly.

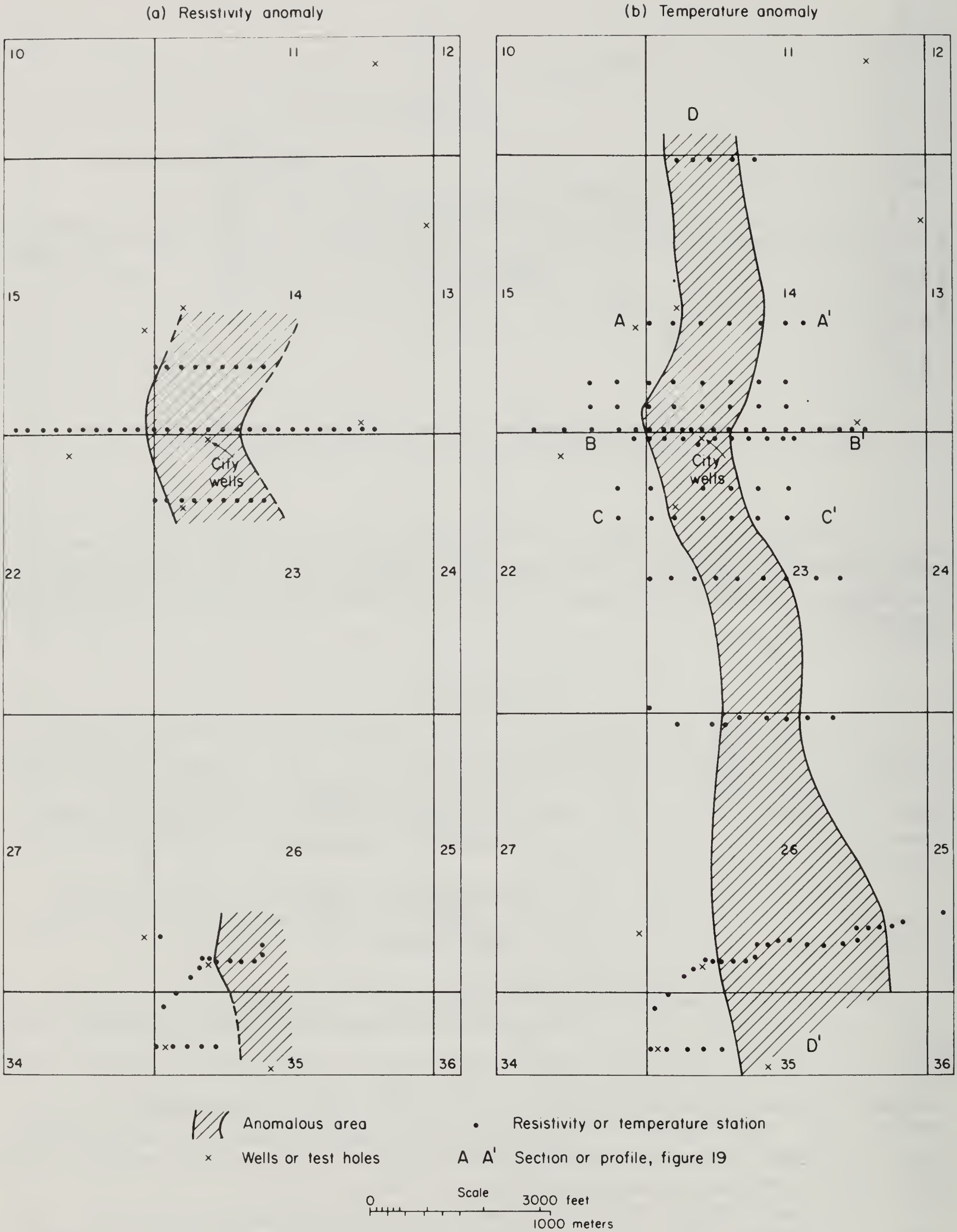

Figure 18 - (a) Resistivity (apparent) anomaly at 30-foot electrode spacing (from Buhle, 1938), and
(b) Temperature anomaly 18 inches below land surface near Mazon, T. 32 N., R. 7 E., Grundy County, Illinois.

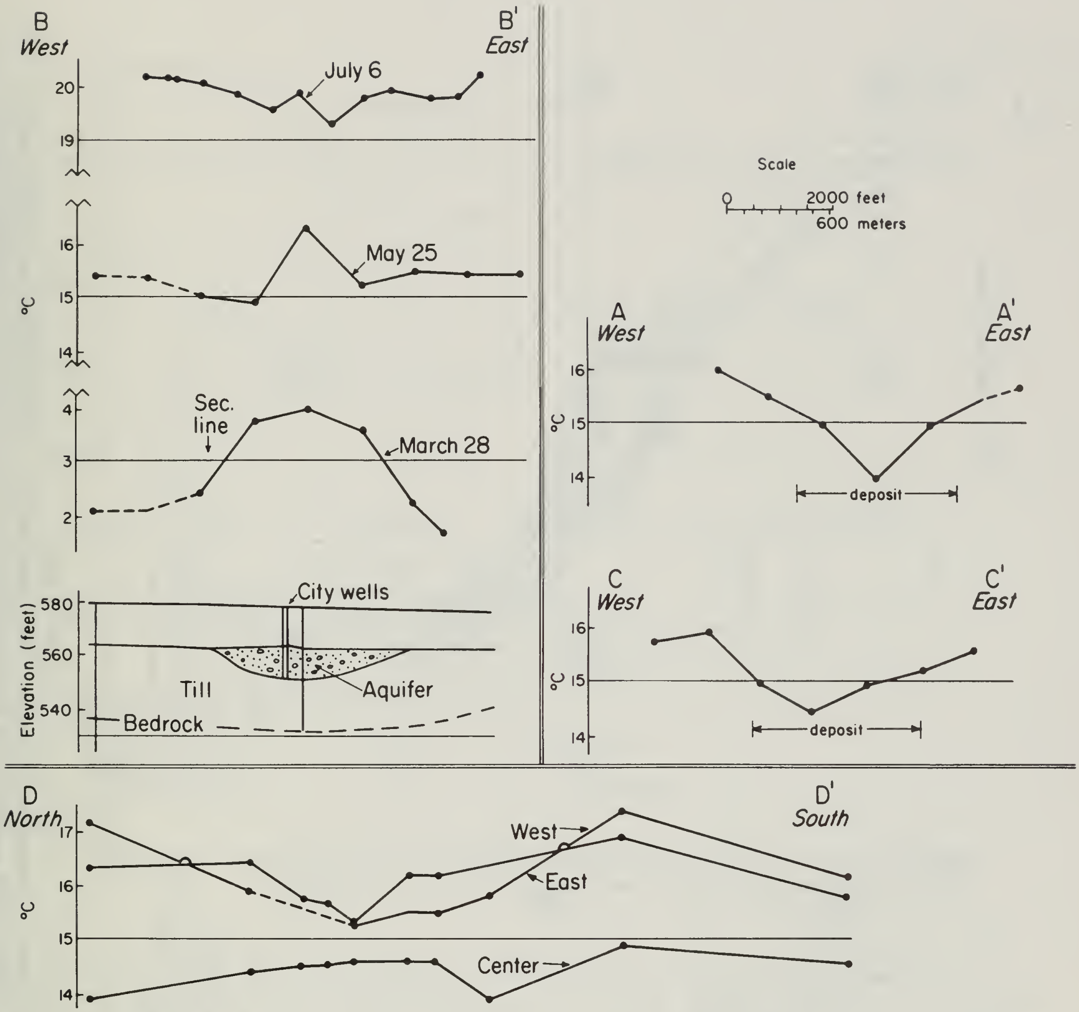

Figure 19 - Temperature profiles (summer and winter) along A-A', B-B', and C-C' (fig. 18), a geologic cross section along B-B', and longitudinal temperature profile along D-D' through the western and eastern edges and center of the deposit near Mazon.

Mulberry Grove

Mulberry Grove is a small town on the Illinoian till plain 55 miles northeast of East St. Louis in south-central Illinois (fig. 10). The deposit is in Hurricane Creek (not the same Hurricane Creek as the first example) east of the town (Pryor, 1955). The creek has cut through the till and into the impermeable Pennsylvanian age shale and sandstone, over which is a maximum of about 30 feet (9 meters) of alluvial material in the valley.

The aquifer, with a maximum thickness of 11 feet (3.3 meters), varies rapidly in character from fine-grained sand to coarse-grained sand and gravel

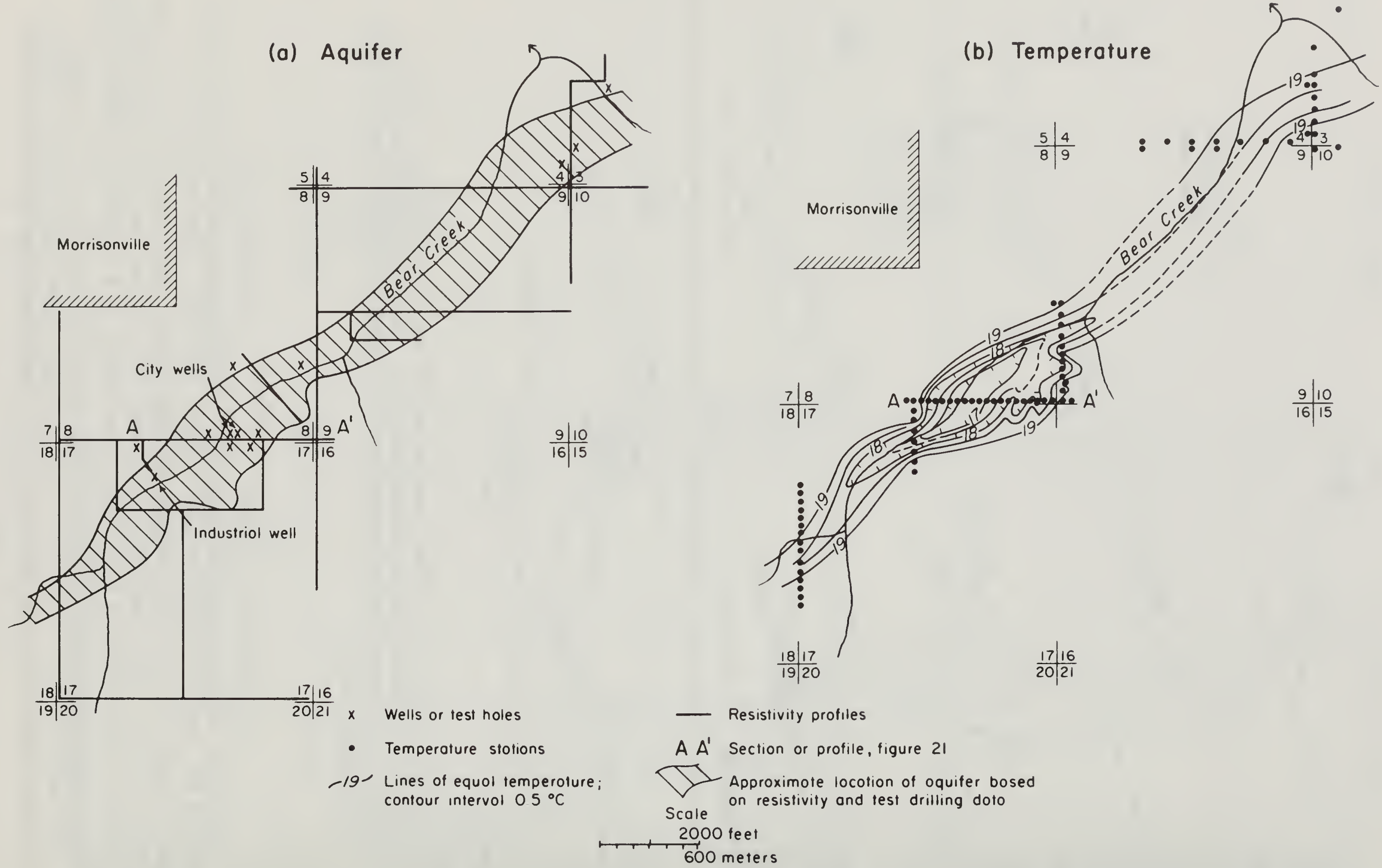

Figure 20 - (a) Location of the aquifer based on resistivity (Buhle, 1942) and geologic data (Cartwright, 1963), and (b) Temperature 18 inches below land surface near Morrisonville, T. 11 N., R. 3 W., Christian County, Illinois.

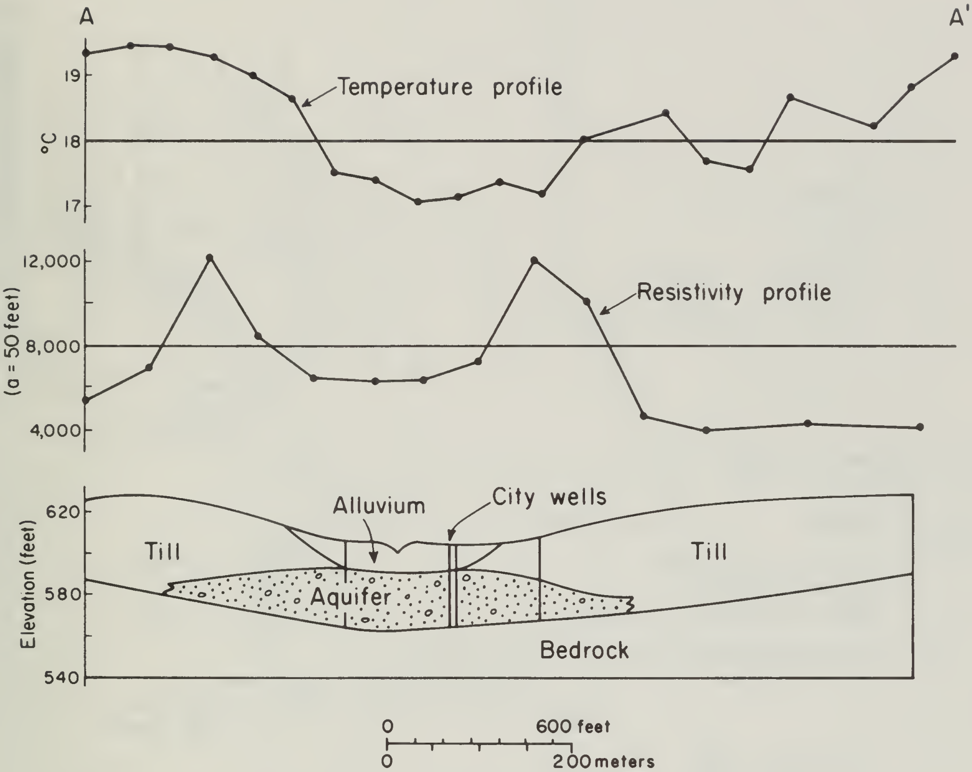

Figure 21 - Temperature and resistivity (apparent) profiles and geologic cross section along A-A' (fig. 20) near Morrisonville.

(Cartwright, 1963). It is generally underlain by the bedrock and overlain by silty clay alluvium. The deposit appears to be very narrow, about 100 feet (30 meters) wide or less, and sinuous in pattern, probably being point bar on stream channel deposits. It is also possible that the deposit may be a series of discontinuous sand bars.

The Mulberry Grove deposit (fig. 22a) is the only one on which the temperature survey was not entirely successful, as was also the case with the resistivity surveys of the site. Both summer and winter surveys were made (fig. 22b, c). An anomaly of as much as 1.5° C was observed in both summer and winter. Some of the same warming effect around pumping wells was observed as at Mazon.

The profiles at the line of cross section (fig. 23) show an anomaly of 1.0° C in the summer and 0.8° C in the winter over the deposit. The western side of the deposit is not as sharply defined as the eastern side.

Mt. Pulaski

Mt. Pulaski is a small town on the Illinoian till plain 25 miles northeast of Springfield (fig. 10). The well field is located $2\frac{1}{2}$ miles north of town in the valley of Salt Creek (Buhle, 1959), which carried outwash from the Wisconsinan ice at its maximum extent located about 12 miles to the east.

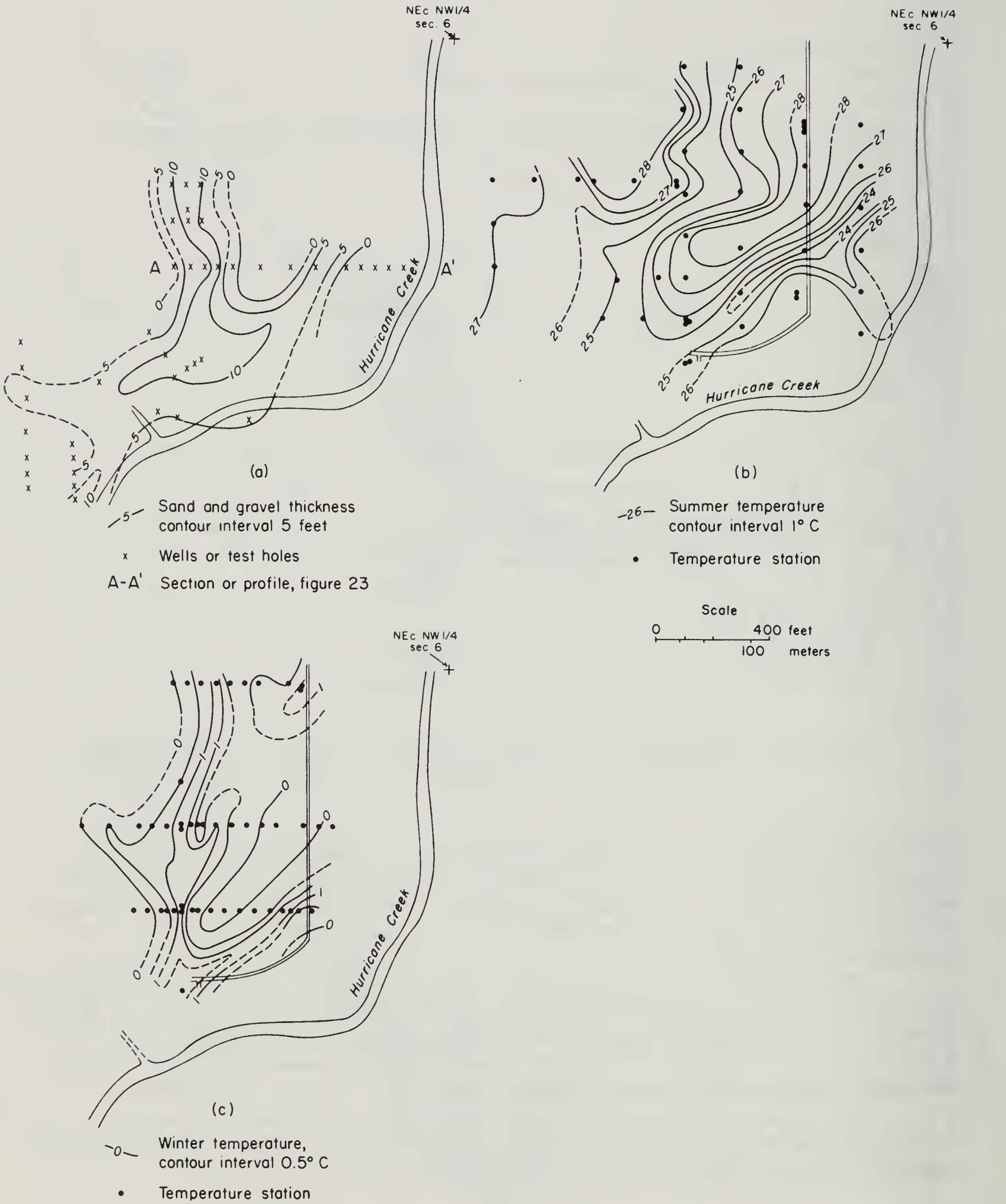

Figure 22 - (a) Thickness of sand,
(b) Summer temperature, and
(c) Winter temperature 18 inches below land surface near Mulberry Grove, sec. 6, T. 5 N., R. 1 W., Fayette County, Illinois.

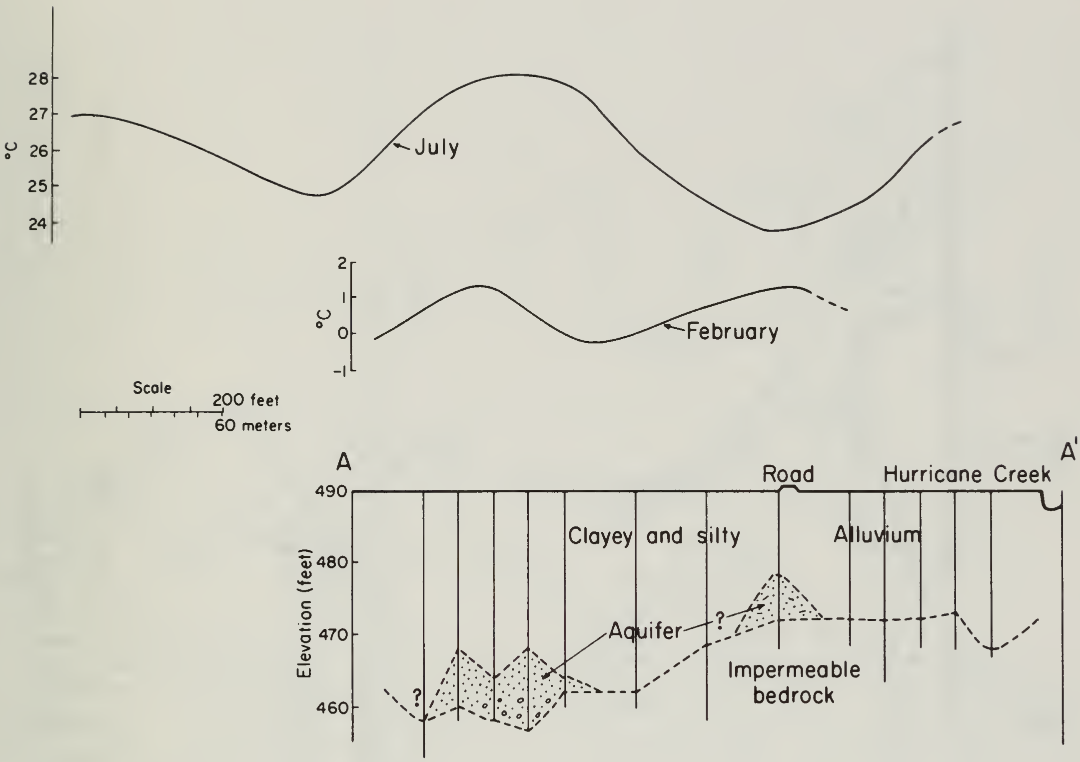

Figure 23 - Temperature profiles (summer and winter) and geologic cross section along A-A' (fig. 22) near Mulberry Grove.

The aquifer lies at a depth of 6 to 10 feet (2 to 3 meters) below the surface, and ranges up to about 30 feet (9 meters) thick. The deposit consists mostly of clean, medium- to coarse-grained sand commonly with fine-grained gravel in the upper half of the aquifer. Sand and gravel is present to some extent over the entire valley.

The Mt. Pulaski survey is the only temperature survey made entirely in the winter months (February-March 1967). The resistivity map (fig. 24a), thermal map (fig. 24b), and cross section (fig. 25) show close agreement. The maximum anomaly is about 3° C, although it is generally slightly less than 2° C. There is an additional 1° C temperature difference between the upland and valley areas. The boundary of the principal water-yielding area is drawn on the 10,000 ohm-cm apparent resistivity at the 40 feet spacing of the electrodes; this matches very well with the area of maximum temperature anomaly.

Spring Valley

Spring Valley is in Bureau County on the bluff north of the Illinois River, about 90 miles west-southwest of Chicago (fig. 10). The city draws water from three ground-water sources, one of which is a shallow sand and gravel deposit on the western edge of the city.

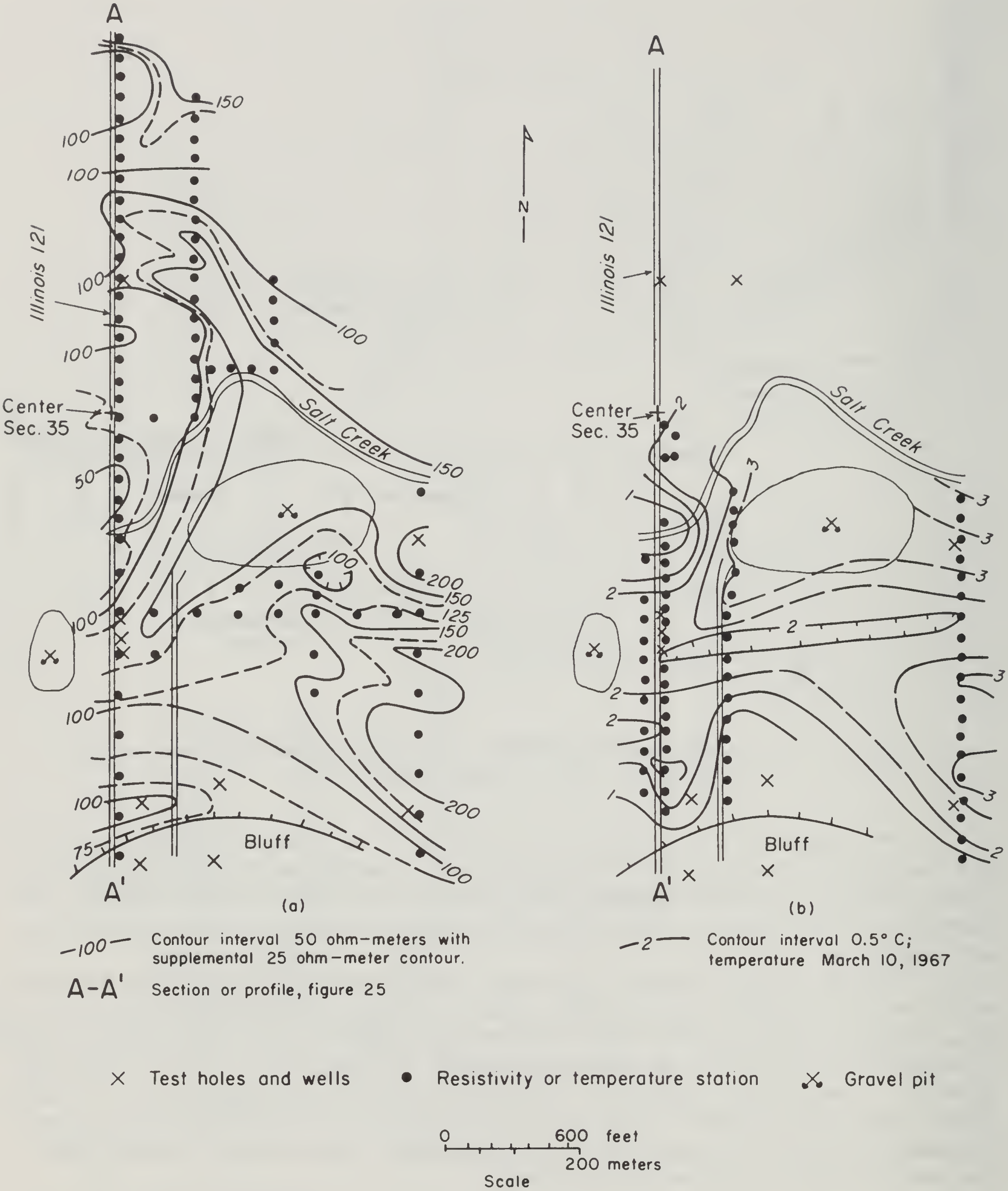

Figure 24 - (a) Resistivity (apparent) at a 40-foot electrode spacing, and (b) Temperature 18 inches below land surface near Mt. Pulaski, sec. 35, T. 19 N., R. 2 W., Logan County, Illinois.

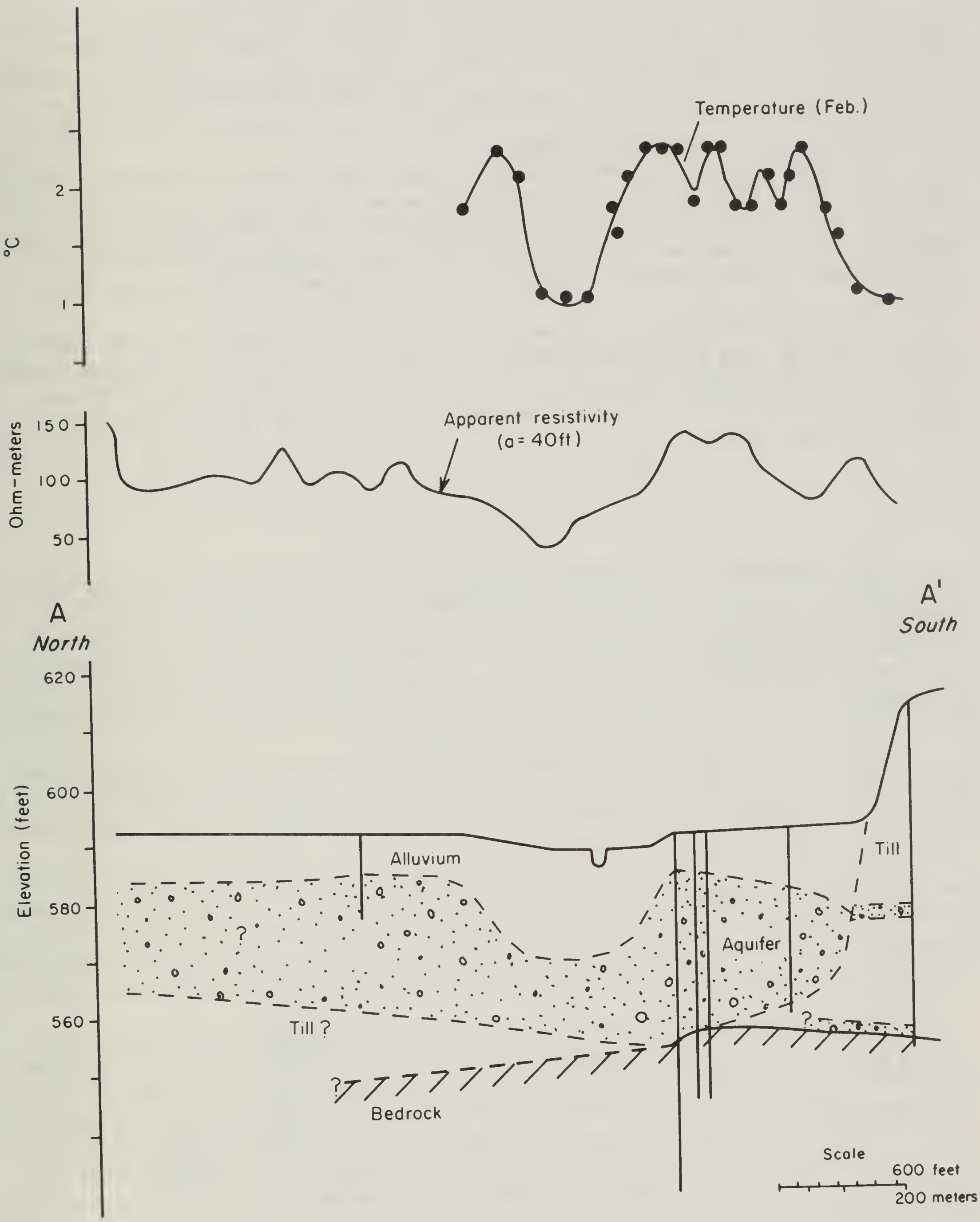

Figure 25 - Temperature and resistivity (apparent) profiles and geologic cross section along A-A' (fig. 24) near Mt. Pulaski.

The area is one of Wisconsinan ground moraine with Illinoian till exposed in low ground (Cady, 1919). The aquifer lies under the sloping ground where most of the Wisconsinan till has been removed. The deposit itself lies at a depth of 20 to 40 feet (6 to 12 meters), depending upon surface elevations. It has a maximum thickness of about 15 feet (4.5 meters) and is overlain by glacial till and underlain by Pennsylvanian shales.

The deposit is fairly well outlined by resistivity surveys and drilling (Buhle, 1945) (fig. 26a). The temperature survey, made late in March, was not very extensive. It shows a warm summer type anomaly (fig. 26b) instead of a winter type anomaly, although the soil is still rather cool. The temperature profile (fig. 27) shows a rather erratic pattern, suggesting the system had not yet entirely stabilized.

The early change to a summer type anomaly is attributed to a warm March, during which the soil had generally begun to warm up; the area not underlain by the aquifer generally warmed faster than the areas over the aquifer. Some frost was still in the ground over the aquifer, and large variation was due to different soil colors and soil cover.

CONCLUSIONS

The theoretical consideration of some of the thermal properties of shallow alluvial and glacial aquifers and the properties of the overburden suggests that aquifers can be detected at the land surface by anomalies in the soil temperature. The theory is reasonably well substantiated by field studies of known aquifers. Calculations of depth to the top of the aquifer are estimates at best by either method proposed (size of the anomaly or two-point method), as many of the data needed are based on an estimate of the parameters involved.

The field studies show that large areas can be covered in a relatively short time. However, long time lapses (in excess of three or four weeks) cannot be compensated for by simple measurements of soil temperature drift; changes in the whole system must then be taken into account.

Temperature variations caused by changes in vegetation and shade were the two most difficult problems faced in the field. Variations in soil temperature caused by these factors were as large as, and sometimes larger than, the anomaly due to the aquifer. Buried pipelines or other conduits may also cause problems, and are not as readily identified in the field. By careful field work, these problems generally can be reduced.

Rain can also reduce or completely erase the soil temperature anomaly resulting from a shallow buried aquifer. The downward movement of water of uniform temperature will erase the anomaly temporarily, or reduce it to a size that is almost indistinguishable. This is partly dependent on the permeability of the soil and the amount of rain; generally the more permeable the soil, the more easily the anomaly is lost by infiltration of rain water, and the less precipitation necessary to erase the anomaly. Conversely, areas of ground-water discharge can produce an anomaly similar to that expected from an aquifer, when no aquifer is present.

Frost also is a problem in the winter because the formation of ice crystals acts to hold the soil temperature close to the freezing point of water. This problem is easily overcome by measuring the soil temperatures well below the frost zone (at least 25 cm and preferably 50 cm below).

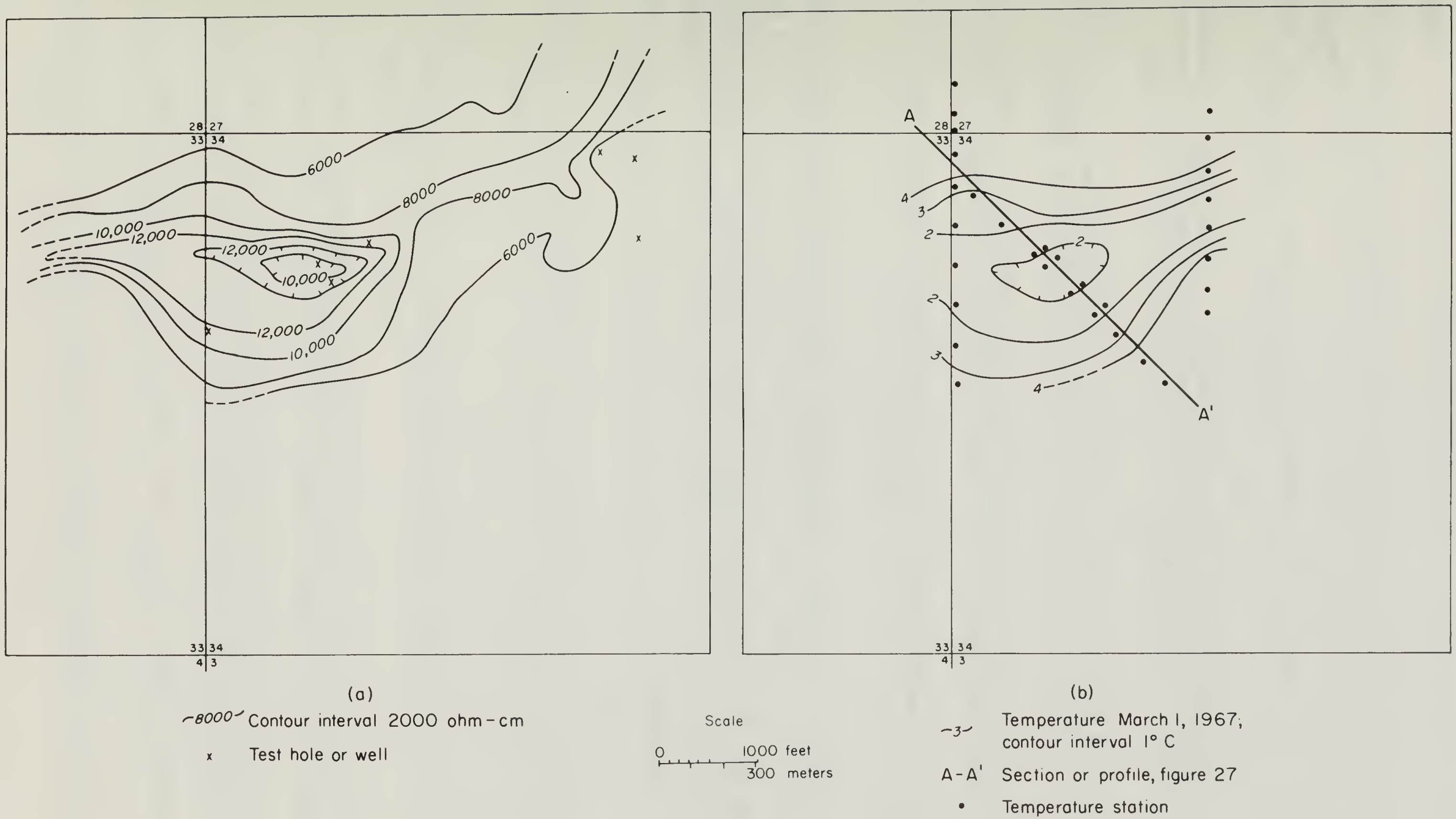

Figure 26 - (a) Resistivity (apparent) at 50-foot electrode spacing, and
(b) Temperature 18 inches below land surface at Spring Valley, sec. 33 and 34, T. 16 N., R. 11 E., Bureau County, Illinois.

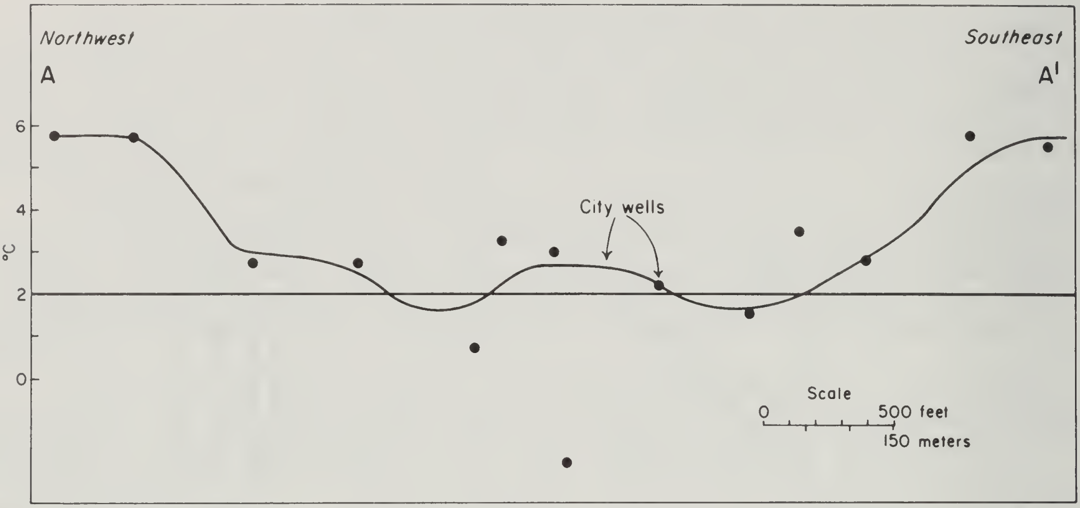

Figure 27 - Temperature profiles across the aquifer at Spring Valley along A-A' (fig. 26).

Nevertheless, temperature exploration offers considerable promise as an inexpensive and fairly reliable means of locating shallow linear glacial and alluvial aquifers during the winter and summer seasons. Information about depth and character of the aquifer by temperature methods is not very reliable. Exact information as to depth, thickness, and water-yielding character are probably best obtained by other means of exploration.

REFERENCES

Birch, Francis, 1942, Thermal conductivity and diffusivity, in Handbook of physical constants: Geol. Soc. America Spec. Paper 36, p. 243-266.

Birman, J. H., 1965, Geothermal prospecting: U. S. Patent 3,217,550, c. 11/16/65, f. 6/11/62.

Bredehoeft, J. D., and I. S. Papadopulos, 1965, Rates of vertical ground-water movement estimated from the earth; thermal profile: Water Resources Res., v. 1, no. 2, p. 325-328.

Buhle, M. B., 1938, Results of an electrical earth resistivity survey at Mazon, Illinois: Illinois Geol. Survey, open file report, unpubl.

______, 1942, An electrical earth resistivity survey in the vicintiy of Morrisonville, Christian County, Illinois: Illinois Geol. Survey, open file report, unpubl.

______, 1945, Results of additional electrical earth resistivity survey at Spring Valley, Illinois: Illinois Geol. Survey, open file report, unpubl.

______, 1953, Earth resistivity in ground-water studies in Illinois: AIME Trans., v. 34, no. 2, p. 395-399.

______, 1959, Supplemental electrical earth resistivity survey for Mt. Pulaski, Logan County, Illinois: Illinois Geol. Survey, open file report, unpubl.

Cady, G. H., 1919, Geology and mineral resources of the Hennepin and LaSalle Quadrangles: Illinois Geol. Survey Bull. 37, 136 p.

Carr, D. D., and R. E. Blakely, 1966, Temperature variation of a depth of 30 cm in clay till and outwash sand and gravel: 4th Symposium on Remote Sensing of Environment Proc., Univ. Michigan, Ann Arbor, p. 203-214.

Cartwright, Keros, 1968, Geologic report on the ground-water conditions in the vicinity of Morrisonville, Christian County, Illinois: Illinois Geol. Survey, open file report, unpubl.

______, 1963, Supplemental geologic report for the city of Mulberry Grove, Bond County, Illinois: Illinois Geol. Survey, open file report, unpubl.

______, 1966, Thermal prospecting for shallow glacial and alluvial aquifers in Illinois (abs.): Program, Ann. Meeting Geol. Soc. America, p. 36.

______, 1968, Thermal prospecting for ground water: Water Resources Research, v. 4, no. 2, p. 395-401.

Changnon, S. A., Jr., 1959, Summary of weather conditions at Champaign-Urbana, Illinois: Illinois Water Survey Bull. 47, 95 p.

Clark, S. P., Jr., 1966, Thermal conductivity, in Handbook of physical constants, rev. ed., Geol. Soc. America Mem. 97, p. 459-482.

Dobrin, M. B., 1952, Introduction to geophysical prospecting: McGraw-Hill, New York, 435 p.

Emery, K. O., 1942, Preliminary report on electrical earth resistivity survey for Oak and Sangamon Ordnance plants near Illiopolis, Illinois: Illinois Geol. Survey, open file report, unpubl.

Flucker, B. J., 1958, Soil temperatures: Soil Sci., v. 86, no. 1, p. 35-46.

Guyod, Hubert, 1946, Temperature well logging; reprint of eight articles that appeared in Oil Weekly, assembled by Halliburton Oil Well Cementing Co.

Handbook of Chemistry and Physics, 1967, 48th edition, C. D. Hodgman, ed., Chemical Rubber Publ. Co., Cleveland, Ohio, p. E-3.

Hanson, Ross, 1950, Public ground-water supplies in Illinois: Illinois Water Survey Bull. 40, 1060 p.

_______, 1958, New public ground-water supplies in Illinois, 1950-1957: Illinois Water Survey Bull. 40, Supp. I, 260 p.

_______, 1961, Addition to public ground-water supplies in Illinois: Illinois Water Survey Bull. 40, Supp. II, 656 p.

Ingersoll, L. R., O. J. Zobel, and A. C. Ingersoll, 1954, Heat conduction, with engineering, geological, and other applications (rev. ed.): Univ. Wisconsin Press, Madison, 325 p.

Jaeger, J. C., 1965, Application of the theory of heat conduction to geothermal measurements, in Terrestrial heat flow: Am. Geophys. Union Geophys. Mon. Ser., no. 8, p. 7-23.

Kintzinger, P. R., 1956, Geothermal survey of hot ground near Lordsburg, New Mexico: Science, v. 124, p. 629.

Loofbourow, R. L., 1966, Depth and rock temperature: Mining Eng., v. 18, no. 1, p. 82-87.

Lovering, T. S., and H. D. Goode, 1963, Measuring geothermal gradients in drill holes less than 60 feet deep, East Tintic District, Utah: U. S. Geol. Survey Bull. 1172, 48 p.

McGinnis, L. D., 1968, Glaciation as a possible cause of mineral deposition (abs.): Am. Geophys. Union Proc., v. 49, no. 2, p. 356.

McGinnis, L. D., J. P. Kempton, and P. C. Heigold, 1963, Relationship of gravity anomalies to a drift-filled bedrock valley system in northern Illinois: Illinois Geol. Survey Circ. 354, 23 p.

Misener, A. D., L. G. D. Thompson, and R. J. Uffen, 1951, Terrestrial heat flow in Ontario and Quebec: Am. Geophys. Union Trans., v. 32, no. 5, p. 729-737.

National Research Council, 1927, International critical tables of numerical data, physics, chemistry, and technology: McGraw-Hill, New York, v. 2, 616 p.

Penrod, E. B., J. M. Elliott, and W. K. Brown, 1960, Soil temperature measurements at Lexington, Kentucky, from 1952-1956: Kentucky Acad. Sci. Trans., v. 21, no. 3, p. 49-60.

Penrod, E. B., W. W. Walton, and D. V. Terrell, 1958, A method to describe soil temperature variation: ASCE Proc., v. 84, pt. 1, no. SM-1, p. 1537-1 to -21.

Pryor, W. A., 1955, Report on a supplemental electrical earth resistivity survey for the Village of Mulberry Grove, Bond County, Illinois: Illinois Geol. Survey, open file report, unpubl.

_______, 1956, Quality of ground water estimated from electric resistivity logs: Illinois Geol. Survey Circ. 215, 15 p.

Radio and Electronics, Feb. 1963, Build the temp—all, p. 45-47.

Spicer, H. C., 1942, Observed temperatures in the earth's crust, in Handbook of physical constants: Geol. Soc. America Spec. Paper 36, p. 279-292.

Stallman, R. W., 1965, Steady one-dimensional fluid flow in a semi-infinite porous medium with sinusoidal surface temperature: Jour. Geophys. Res., v. 70, no. 12, p. 2821-2827.

Strangway, O. W., and R. C. Holmer, 1966, The search for ore deposits using thermal radiation: Geophysics, v. 31, no. 1, p. 225-242.

Suter, Max, R. E. Bergstrom, H. F. Smith, G. H. Emrich, W. C. Walton, and T. E. Larson, 1959, Preliminary report on ground-water resources of the Chicago region, Illinois: Illinois Geol. Survey and Illinois Water Survey Coop. Ground-Water Rept. 1, 89 p.

Van Orstrand, C. E., 1940, Additional evidence on the relation of temperature to structure in the Salt Creek Oil Field, Wyoming: Geophysics, v. 5, p. 47-56.

Walker, W. H., R. E. Bergstrom, and W. C. Walton, 1965, Preliminary report on the ground-water resources of the Havana region in west-central Illinois: Illinois Geol. Survey and Illinois Water Survey Coop. Ground-Water Rept. 3, 61 p.

Illinois State Geological Survey Circular 433
41 p., 27 figs., 4 tables, 2100 cop., 1968
Urbana, Illinois 61801

Printed by Authority of State of Illinois, Ch. 127, IRS, Par. 58.25.

ILLINOIS STATE GEOLOGICAL SURVEY Urbana, Illinois

FULL TIME STAFF
October 1, 1968

JOHN C. FRYE, Ph.D., D.Sc., Chief
Hubert E. Risser, Ph.D., Assistant Chief

R. J. Helfinstine, M.S., Administrative Engineer
G. R. Eadie, M.S., E.M., Asst. Administrative Engineer

Velda A. Millard, Fiscal Assistant to the Chief
Helen E. McMorris, Secretary to the Chief

GEOLOGICAL GROUP

Jack A. Simon, M.S., Principal Geologist
M. L. Thompson, Ph.D., Principal Research Geologist
Frances H. Alsterlund, A.B., Research Assistant

COAL

M. E. Hopkins, Ph.D., Geologist and Acting Head
William H. Smith, M.S., Geologist
Kenneth E. Clegg, M.S., Associate Geologist
Heinz H. Damberger, D.Sc., Associate Geologist
Harold J. Gluskoter, Ph.D., Associate Geologist
Russel A. Peppers, Ph.D., Associate Geologist
John A. Bell, Ph.D., Assistant Geologist

STRATIGRAPHY AND AREAL GEOLOGY

H. B. Willman, Ph.D., Geologist and Head
Elwood Atherton, Ph.D., Geologist
T. C. Buschbach, Ph.D., Geologist
Charles Collinson, Ph.D., Geologist
Herbert D. Glass, Ph.D., Geologist
Lois S. Kent, Ph.D., Associate Geologist
Jerry A. Lineback, Ph.D., Associate Geologist
Alan M. Jacobs, Ph.D., Assistant Geologist
Susan R. Avcin, B.A., Research Assistant

ENGINEERING GEOLOGY AND TOPOGRAPHIC MAPPING

W. Calhoun Smith, Ph.D., Geologist in charge
Paul B. DuMontelle, M.S., Assistant Geologist
Patricia M. Moran, B.A., Research Assistant

GEOLOGICAL RECORDS

Vivian Gordon, Head
Hannah Kistler, Supervisory Technical Assistant
Constance Armstrong, Technical Assistant
Connie L. Maske, B.A., Technical Assistant
Elizabeth Speer, Technical Assistant
Bonnie B. Sullivan, Technical Assistant
Rebecca J. Veenstra, Technical Assistant
Margaret J. Weatherhead, Technical Assistant

GEOLOGICAL SAMPLES LIBRARY

Robert W. Frame, Superintendent
J. Stanton Bonwell, Technical Assistant
Eugene W. Meier, Technical Assistant
Linda D. Rentfrow, Clerk-Typist II

GROUND-WATER GEOLOGY AND GEOPHYSICAL EXPLORATION

Robert E. Bergstrom, Ph.D., Geologist and Head
Merlyn B. Buhle, M.S., Geologist
George M. Hughes, Ph.D., Associate Geologist
John P. Kempton, Ph.D., Associate Geologist
Keros Cartwright, M.S., Assistant Geologist
Manoutchehr Heidari, M.S., Assistant Engineer
Paul C. Heigold, M.S., Assistant Geophysicist
Jean I. Larsen, M.A., Assistant Geologist
Murray R. McComas, M.S., Assistant Geologist
Kemal Piskin, M.S., Assistant Geologist
Frank B. Sherman, Jr., M.S., Assistant Geologist
Shirley A. Masters, B.S., Research Assistant
Verena M. Colvin, Technical Assistant
Daniel E. McMeen, B.A., Technical Assistant

OIL AND GAS

Donald C. Bond, Ph.D., Head
Lindell H. Van Dyke, M.S., Geologist
Thomas F. Lawry, B.S., Associate Petrol. Engineer
R. F. Mast, M.S., Associate Petrol. Engineer
Wayne F. Meents, Associate Geological Engineer
Hubert M. Bristol, M.S., Assistant Geologist
Richard H. Howard, M.S., Assistant Geologist
David L. Stevenson, M.S., Assistant Geologist
Jacob Van Den Berg, M.S., Assistant Geologist
Albert L. Meyers, B.S., Research Assistant

INDUSTRIAL MINERALS

James C. Bradbury, Ph.D., Geologist and Head
James W. Baxter, Ph. D., Associate Geologist
Richard D. Harvey, Ph.D., Associate Geologist
Norman C. Hester, Ph.D., Assistant Geologist

CLAY RESOURCES AND CLAY MINERAL TECHNOLOGY

W. Arthur White, Ph.D., Geologist and Head
Bruce F. Bohor, Ph.D., Associate Geologist

CHEMICAL GROUP

Glenn C. Finger, Ph.D., Principal Chemist
Ruth C. Lynge, Technical Assistant
Thelma J. Chapman, B.A., Technical Assistant

COAL CHEMISTRY

G. Robert Yohe, Ph.D., Chemist and Head

PHYSICAL CHEMISTRY

Josephus Thomas, Jr., Ph.D., Chemist and Head
Robert N. Leamnson, M.S., Assistant Chemist

ORGANIC GEOCHEMISTRY

G. C. Finger, Ph.D., Acting Head
Donald R. Dickerson, Ph.D., Associate Chemist
Richard H. Shiley, M.S., Assistant Chemist
Gilbert L. Tinberg, Technical Assistant

CHEMICAL ENGINEERING

H. W. Jackman, M.S.E., Chemical Engineer and Head
R. J. Helfinstine, M.S., Mechanical Engineer
Henry P. Ehrlinger III, M.S., Assoc. Minerals Engineer
Lee D. Arnold, B.S., Assistant Engineer
W. G. ten Kate, M.S., Geol. D., Assistant Mineralogist
Walter E. Cooper, Technical Assistant
Robert M. Fairfield, Technical Assistant
John P. McClellan, Technical Assistant
Edward A. Schaede, Technical Assistant (on leave)

(Chemical Group continued on next page)

CHEMICAL GROUP (Continued)

ANALYTICAL CHEMISTRY

Neil F. Shimp, Ph.D., Chemist and Head
William J. Armon, M.S., Associate Chemist
Charles W. Beeler, M.A., Associate Chemist
Rodney R. Ruch, Ph.D., Associate Chemist
John A. Schleicher, B.S., Associate Chemist
Larry R. Camp, B.S., Assistant Chemist
David B. Heck, B.S., Assistant Chemist
L. R. Henderson, B.S., Assistant Chemist
Stephen M. Kim, B.A., Assistant Chemist
John K. Kuhn, B.S., Assistant Chemist
Ru-tao Kyi, Ph.D., Assistant Chemist
Sharon L. Olson, B.S., Special Research Assistant
Paul E. Gardner, Technical Assistant
George R. James, Technical Assistant

MINERAL ECONOMICS GROUP

Hubert E. Risser, Ph.D., Principal Mineral Economist
W. L. Busch, A.B., Associate Mineral Economist
Robert L. Major, M.S., Assistant Mineral Economist

ADMINISTRATIVE GROUP

George R. Eadie, M.S., E.M., Administrator

EDUCATIONAL EXTENSION
David L. Reinertsen, A.M., Associate Geologist in charge
George M. Wilson, M.S., Geologist
William E. Cote, M.S., Assistant Geologist
Helen S. Johnston, B.S., Technical Assistant
Myrna M. Killey, B.A., Technical Assistant

PUBLICATIONS
Betty M. Lynch, B.Ed., Technical Editor
Carol A. Brandt, B.A., Technical Editor
Jane E. Busey, B.S., Assistant Technical Editor
Marie L. Martin, Geologic Draftsman
James R. Gilmer, Asst. Geologic Draftsman
Sandra L. Oncken, B.F.A., Asst. Geologic Draftsman
William Dale Farris, Research Associate
Kathy Sue Billingsley, Technical Assistant
Dorothy H. Scoggin, Technical Assistant
Beulah M. Unfer, Technical Assistant

GENERAL SCIENTIFIC INFORMATION
Peggy H. Schroeder, B.A., Research Assistant
Florence J. Partenheimer, Technical Assistant

SPECIAL TECHNICAL SERVICES
Glenn G. Poor, Research Associate (on leave)
Merle Ridgley, Research Associate
Wayne W. Nofftz, Supervisory Technical Assistant
Donovon M. Watkins, Technical Assistant
Mary M. Sullivan, Supervisory Technical Assistant
David B. Cooley, Technical Assistant

FINANCIAL OFFICE
Velda A. Millard, in charge
Marjorie J. Hatch, Clerk IV
Virginia C. Smith, B.S., Account Clerk
Pauline Mitchell, Account Clerk

CLERICAL SERVICES
Nancy J. Hansen, Clerk-Stenographer II
Hazel V. Orr, Clerk-Stenographer II
Mary K. Rosalius, Clerk-Stenographer II
Dorothy M. Spence, Clerk-Stenographer II
Jane C. Washburn, Clerk-Stenographer II
Becky L. Dowds, Clerk-Stenographer I
Magdeline E. Hutchison, Clerk-Stenographer I
Edna M. Yeargin, Clerk-Stenographer I
Sharon K. Zindars, Clerk-Stenographer I
Shirley L. Weatherford, Key Punch Operator II
Pauline F. Tate, Clerk-Typist II
JoAnn L. Hayn, Clerk-Typist I

TECHNICAL RECORDS
Berenice Reed, Supervisory Technical Assistant
Miriam Hatch, Technical Assistant
Hester L. Nesmith, B.S., Technical Assistant

LIBRARY
Lieselotte F. Haak, Geological Librarian (on leave)
Ann M. Sokan, M.A., Acting Geol. Librarian
Jo Ann Boeke, Technical Assistant

EMERITI
M. M. Leighton, Ph.D., D.Sc., Chief, Emeritus
J. S. Machin, Ph.D., Principal Chemist, Emeritus
O. W. Rees, Ph.D., Prin. Research Chemist, Emeritus
W. H. Voskuil, Ph.D., Prin. Mineral Economist, Emeritus
G. H. Cady, Ph.D., Senior Geologist, Emeritus
A. H. Bell, Ph.D., Geologist, Emeritus
George E. Ekblaw, Ph.D., Geologist, Emeritus
J. E. Lamar, B.S., Geologist, Emeritus
R. J. Piersol, Ph.D., Physicist, Emeritus
L. D. McVicker, B.S., Chemist, Emeritus
Enid Townley, M.S., Geologist, Emerita
Lester L. Whiting, M.S., Geologist, Emeritus
Juanita Witters, M.S., Physicist, Emerita
B. J. Greenwood, B.S., Mechanical Engineer, Emeritus

RESEARCH AFFILIATES AND CONSULTANTS
Richard C. Anderson, Ph.D., Augustana College
W. F. Bradley, Ph.D., University of Texas
Donald L. Graf, Ph.D., University of Minnesota
Ralph E. Grim, Ph.D., University of Illinois
S. E. Harris, Jr., Ph.D., Southern Illinois University
Lyle D. McGinnis, Ph.D., Northern Illinois University
I. Edgar Odom, Ph.D., Northern Illinois University
T. K. Searight, Ph.D., Illinois State University
Harold R. Wanless, Ph.D., University of Illinois
George W. White, Ph.D., University of Illinois

Topographic mapping in cooperation with the United States Geological Survey.

CIRCULAR 433

ILLINOIS STATE GEOLOGICAL SURVEY

URBANA